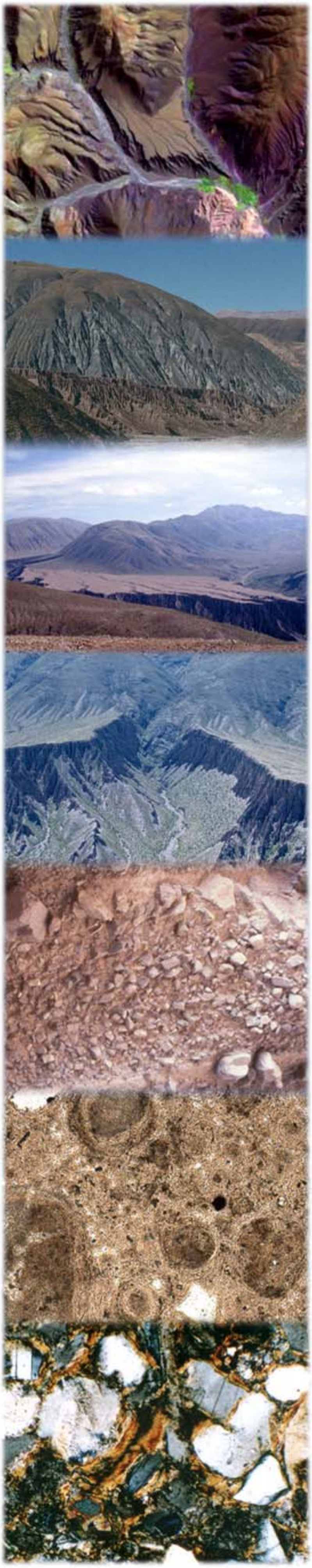

Diplomarbeit

Geographisches Institut
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The Quebrada de Purmamarca, Jujuy, NW-Argentina:

Landscape Evolution and Morphodynamics in the Semi-Arid Andes

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November 2002

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THE QUEBRADA DE PURMAMARCA, JUJUY, NW-ARGENTINA: LANDSCAPE EVOLUTION AND MORPHODYNAMICS IN THE SEMI-ARID ANDES

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THE QUEBRADA DE PURMAMARCA, JUJUY, NW-ARGENTINA: LANDSCAPE EVOLUTION AND MORPHODYNAMICS IN THE SEMI-ARID ANDES

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ABBREVIATIONS

IGM	Instituto Geográfico Militar
SEGEMAR	Servicio Geológico Minero Argentino
ITGE	Instituto Tecnológico Geominero de España
FAO	Food and Agriculture Organization of the United Nations
GSA	Geological Society of America
USGS	United States Geological Survey
EDC	Eros Data Center
STN	Secretaría de Tursimo de la Nación
GIS	Geographical Information System
NW-Argentina	Northwestern Argentina
ITCZ	Inner-Tropical Convergence Zone
NCDC	National Climatic Data Center
ENSO	El Niño Southern Oscillation
LGM	Late Glacial Maximum
ka	1,000 years
Ma	1,000,000 years
AD	Anno Domini
BP	before present (1950)
e.g.	for example
i.e.	that is
m.a.s.l.	Meters above sea level
km	Kilometers
m	Meters
cm	Centimeters
mm	Millimeters

STUDY OUTLINE

As implied by the title, the goal of this study is the *reconstruction of a landscape evolution with particular attention to geomorphology.* CHAPTER 1 provides a short background to this project, points out its importance and sets the theoretical frame for this task. The geographical overview in CHAPTER 2 gives relevant information about the study area and forms the base for the following chapters. CHAPTER 3 outlines the methodical approach of the study as a whole. It comments the involved procedures, emphasizes the role of the geomorphological map, an essential output of this thesis, and draws attention to occurred problems.

CHAPTER 4 step by step describes the observed landforms of the study area, proceeding from larger and older to smaller and younger forms. Each form is interpreted concerning processes of their formation. These first results are then discussed regarding their importance within the overall system of landforms and compared to regional literature where necessary. Aside from geomorphological observations, sedimentological and pedological data from the study area are presented. Particularly in context with the geomorphological results these data contribute further information about the landscape evolution.

CHAPTER 5 presents and discusses all the essential results from the previous chapter. By assigning relative ages to the individual landforms and discussing relevant processes of their formation, eventually the landscape evolution of the study area is unravelled and integrated into a regional frame. Finally, CHAPTER 6 draws conclusions from this study.

1. INTRODUCTION

1.1. BACKGROUND TO THIS STUDY

In November 2002, this project of landscape evolution and geomorphology in NW-Argentina looks back on a history of more than three years. The idea to make it the topic of my master thesis was born during three months of internship at the UNSa (University of Salta, Argentina) from July to September 1999. Besides getting familiar with the Spanish language and South American culture, I got the opportunity to learn about the geology and geomorphology of the semi-arid Andes of NW-Argentina, a scenery rather different from familiar European or North American landscapes.
Personal friendship as well as ongoing academic exchange have vividly continued up to the present day and have motivated my decision to choose a regionally rather remote topic for my thesis. However, this decision required another three months of field work in Argentina, which were completed during spring 2001. Most laboratory, mapping, graphical and analytical work as well as the composition of the text were carried out between February and November 2002.

1.2. LANDSCAPE EVOLUTION AND GEOMORPHOLOGY: CONCEPTS AND SIGNIFICANCE

Landforms are traditionally the object of geomorphological research. They are transitory events in time and space; they emerge, evolve and go by. *Landscape* as a somewhat blurry expression has been geomorphologically defined by most authors as follows (BLOOM 1998, p.11):

> *"Landscapes are surfaces composed of an assemblage of subjectively defined components. Each element of the landscape that can be observed in its entirety, and has consistence of form or regular change of form, is defined as a landform."*

This definition is certainly subjective as far as spatial and temporal scale are concerned. Nevertheless it stresses landforms to be the essential component of any landscape (*landform assemblage*). Besides pure description, geomorphologic research has been focused on the processes forming the landforms and their relation to landform as well as their evolution over time. From this point of view, geomorphology comprises the three aspects of morphography, functional geomorphology or morphodynamics, and historic-genetic geomorphology (AHNERT 1996).

Historic-genetic geomorphology, and therefore *landscape evolution*, will always be based on existing geomorphological components and principles, applying existing knowledge of processes and landforms to landform assemblages from the past. As mentioned above every landscape evolves over time due to the geomorphologic processes shaping them, a fact, which is evidenced by manifold relict landscapes in all parts of the world. Unlike physical and chemical rules, which are valid independent of time (*the concept of uniformitarianism*), geomorphic processes do change over time as they are controlled by a variety of changing factors themselves. This fact certainly limits the concept of uniformitarianism for landscape evolutionary research and constitutes the great danger of pragmatic regional or temporal extrapolations within historic-genetic geomorphological studies (BLOOM 1998). Nevertheless, the recognition of the landscape as a mosaic of present and relict landforms forms the base for further interpretation regarding the causes and mode of change (BÜDEL 1977).

Similarly to other evolutionary theories, the mode of geomorphic change (*evolution*) has been an issue of ongoing discussion. Whether landscapes – and therefore its elements – change abruptly or sequentially largely depends on the applied temporal and spatial resolution. In this context, CHORLEY (1962) introduced the concept of geomorphology as an open system. This system is a system in constant dynamic change, comprising landforms as well as the processes acting upon them. The role of time changes with the temporal scale applied; the system can remain in states of equilibria or it can cross certain thresholds before shifting to a different mode of function. Naturally, the larger the extent of any landform or landform assemblage, the longer it will persist (AHNERT 1996). Thus, the intensity and rates of geomorphic processes acting upon any landform, and therefore the mode of geomorphic change, will always be time and area dependent (WOLMAN AND MILLER 1960, SCHUMM AND LICHTY 1965).
Longer time spans necessarily need to integrate geological and tectonic change (e.g. SUMMERFIELD 2000) and particularly climatic change (e.g. BÜDEL 1977) as controlling and independent variables for geomorphic change, as long as there is evidence for either of them. This is particularly true for the analysis of landscape evolution, where time is typically measured in longer time periods of up to millions of years (BLOOM 1998). Consequently, landscape evolutionary studies also contribute to Quaternary science in general.

Since the emergence of modern age dating possibilities, landscape evolutionary research has become more than the reconstruction of an erosional history of a given landscape, as it now has the abilities to integrate qualitative and quantitative geomorphologic information and allows practicable application of this data to modern geomorphic and environmental research. Research in historic-genetic geomorphology has added valuable information to the understanding of geomorphic processes acting at longer, non-observable timescales in general. In addition, it contributes to the present knowledge about environmental change, both geomorphic and climatic. This is of great interest to every scientist of paleoenvironmental sciences concerned with the reconstruction of times past. But historic-genetic geomorphology also has the duty to direct its findings to the future development of our planet, an issue of local as well as global interest. Having in mind past changes it seems reasonable to create scenarios of future change. Nevertheless, the documentation of the natural world alone is of such relevance to environmental understanding that historic-genetic geomorphology as a part of geoscientific research certainly has the power to close the existing gap between natural science and social concerns (BAKER 1994).

1.3. LANDSCAPE EVOLUTION AND GEOMORPHOLOGY IN NW-ARGENTINA: A REVIEW

So far, most of the existing geoscientific literature of NW-Argentina has been concerned with geological content while the concept of historic-genetic investigations in geomorphology have been underrepresented. A thorough reconnaissance of regional geology commenced in the late 19th century, when studies still had a very descriptive and expeditionary character and researchers generally came from overseas. The first to study the northern provinces of Argentina at a more detailed scale was BRACKEBUSCH, who published his observations in a geological map of Argentina's interior in 1891 (HAUSEN 1923). He was followed by several researchers during the early 20th century whose studies essentially contributed to geologic exploration and reconnaissance.

Geomorphological and landscape evolutionary problems were largely neglected (GONZÁLES DÍAZ 1993), until KEIDEL published his treatise of "*Young fluviatile deposits in the Andes of Argentine*" in 1913, where he introduces first ideas and concepts about the geomorphological evolution of NW-Argentina in relation to Quaternary climatic changes. In this context HAUSEN (1923) includes several of these observations in his geological overview of Salta and Jujuy, particularly mentioning the Quebrada de Purmamarca for its morphological characteristics. An overview of the regional morphology is then given by KÜHN in 1924, including first considerations of present processes and landforms in the study area as well as their evolution. To these thoughts further considerations about the morphology of the study area were added by FOCHLER-HAUKE (1952), CZAIJKA (1958, 1972), VIERS (1967), IGARZABAL (1971) and WERNER (1972, 1984), to name the more commonly known publications on the geomorphology of the region. For the Quebrada de Purmamarca DE FERRARIS (1940) contributed very valuable information in his study on "*Corrimiento de Bloques de Montaña*".

While most geomorphological research remained very descriptive until the 1980's, many universities and research institutes have recently focused their systematic investigations on the Central Andes including NW-Argentina (e.g. SFB 267 OF UNIVERSITIES OF POTSDAM, BERLIN AND GEOFORSCHUNGSZENTRUM POTSDAM; UNIVERSITY OF BAMBERG; UNIVERSITY OF BAYREUTH; UNIVERSITY OF BERN, SWITZERLAND; UNIVERSITY OF BUENOS AIRES, ARGENTINA; NATIONAL UNIVERSITIES OF SALTA AND JUJUY, ARGENTINA; CORNELL UNIVERSITY, ITHACA, USA). In addition, this geomorphologically highly dynamic region has repeatedly made its appearance even in geoscientific textbooks (e.g. BLUME 1994, DURÁN ET AL. 1998).

1.4. AIMS OF THIS STUDY

The aim of this study is to unravel the landscape evolution for the Quebrada de Purmamarca. To fulfil this task, the problem is mainly implemented by combining geomorphological data with pedological and sedimentological data and by deducing a chronological order from their association as well as from absolute age determinations. To some extent, this will allow extrapolation beyond the region and comparison with existing research and to draw a picture of present geomorphic processes in high-mountain drylands.

Keeping in mind the limited number of existing geomorphological and landscape evolutionary research on the region, this study contributes to regional research within the disciplines of dryland geomorphology and Quaternary science. Particularly in combination with additional projects (UNIVERSITY OF POTSDAM), this study may serve as a further piece in the puzzle of the Cenozoic landscape evolution of the region. Although *"it is impossible to quantify the relative importance of climate-controlled processes and structure in the denudation of various tectonic terranes"* (BLOOM 1998, p. 340), qualitative information will add a lot to existing knowledge in geological, geomorphological and paleoclimatic research in a wider sense.
Furthermore, the importance of scientific research for a great number of applications has to be emphasized. In this context, estimating the regional and local risk potential is a significant issue within the field of natural hazard research, particularly for planning and development purposes. Even though sparsely populated, the regional importance of the Quebrada de Purmamarca is enormous. The national road No. 52 constitutes the only direct connection to Chile. In addition a gas pipeline traverses the Quebrada de Purmamarca linking NW-Argentina to Chile's Pacific coast. For the entire Quebrada de Humahuaca several authors have emphasized the need for a better understanding of this highly dynamic environment (CHAYLE AND WAYNE 1995, SOLIS AND OROSCO 1996).

In the widest sense, the study therefore opens the door for further regional research not only in geology, geomorphology, and paleoclimatic sciences, but will hopefully also form the base for the solution of more applied problems of local and regional interest.

2. GEOGRAPHICAL OVERVIEW OF THE STUDY AREA

2.1. GEOGRAPHICAL LOCATION

The study area of this thesis is the Quebrada de Purmamarca, a drainage basin in NW-Argentina approximately 410 km² large. It can be framed loosely by the geographical coordinates *23° 34,2' S, 65° 40' W* (NW corner) and *23° 49,2' S, 65° 27,8' W* (SE corner). Its main settlement, the village of Purmamarca, is situated in the lower reaches of the valley at 23° 44' S, 65° 29' W. The entire area is included in the topographic map of the Republic of Argentina, sheet "Liberator General San Martín" at a scale of 1:250,000 (IGM 1989).

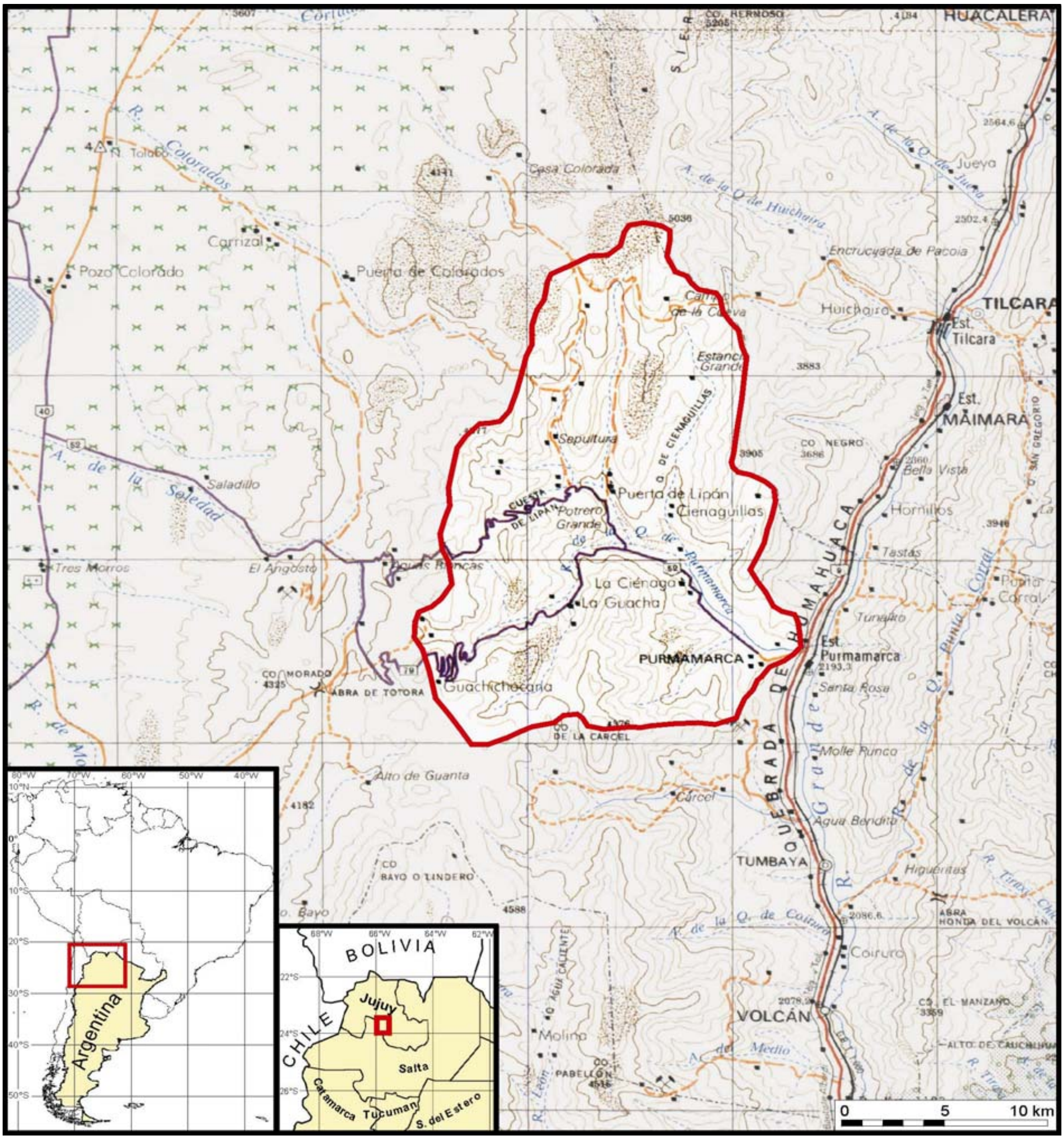


Fig. 1: Topographic overview of the study area (detail from IGM 1989).

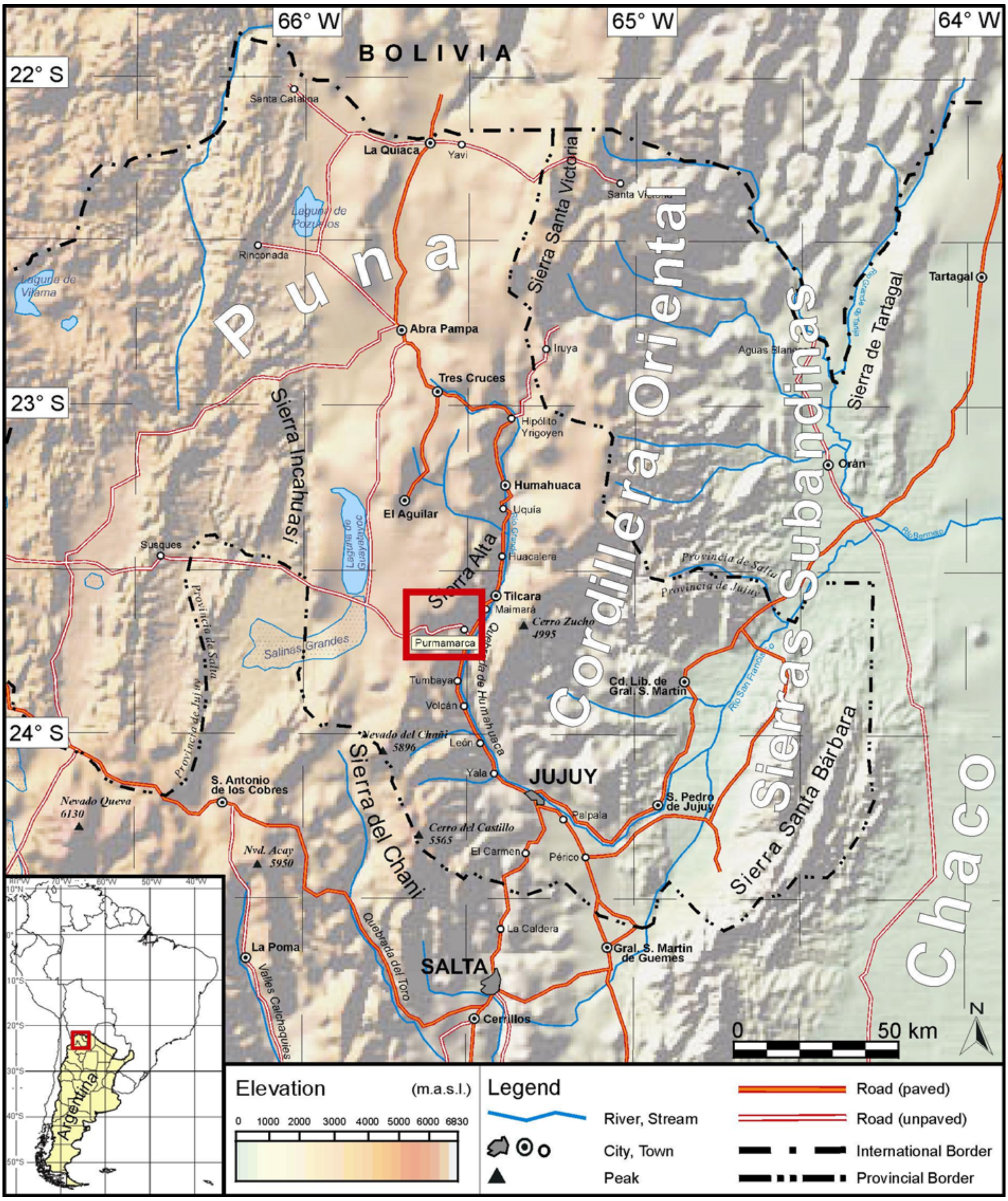


Fig. 2: Topographic and morphological overview of the Jujuy province, NW-Argentina (map composed by author 2002; topographic data from IGM atlas 1999, elevation data from GTOPO30 DEM data, processed and visualized by HEINZ SCHÖFFMANN, DLR).

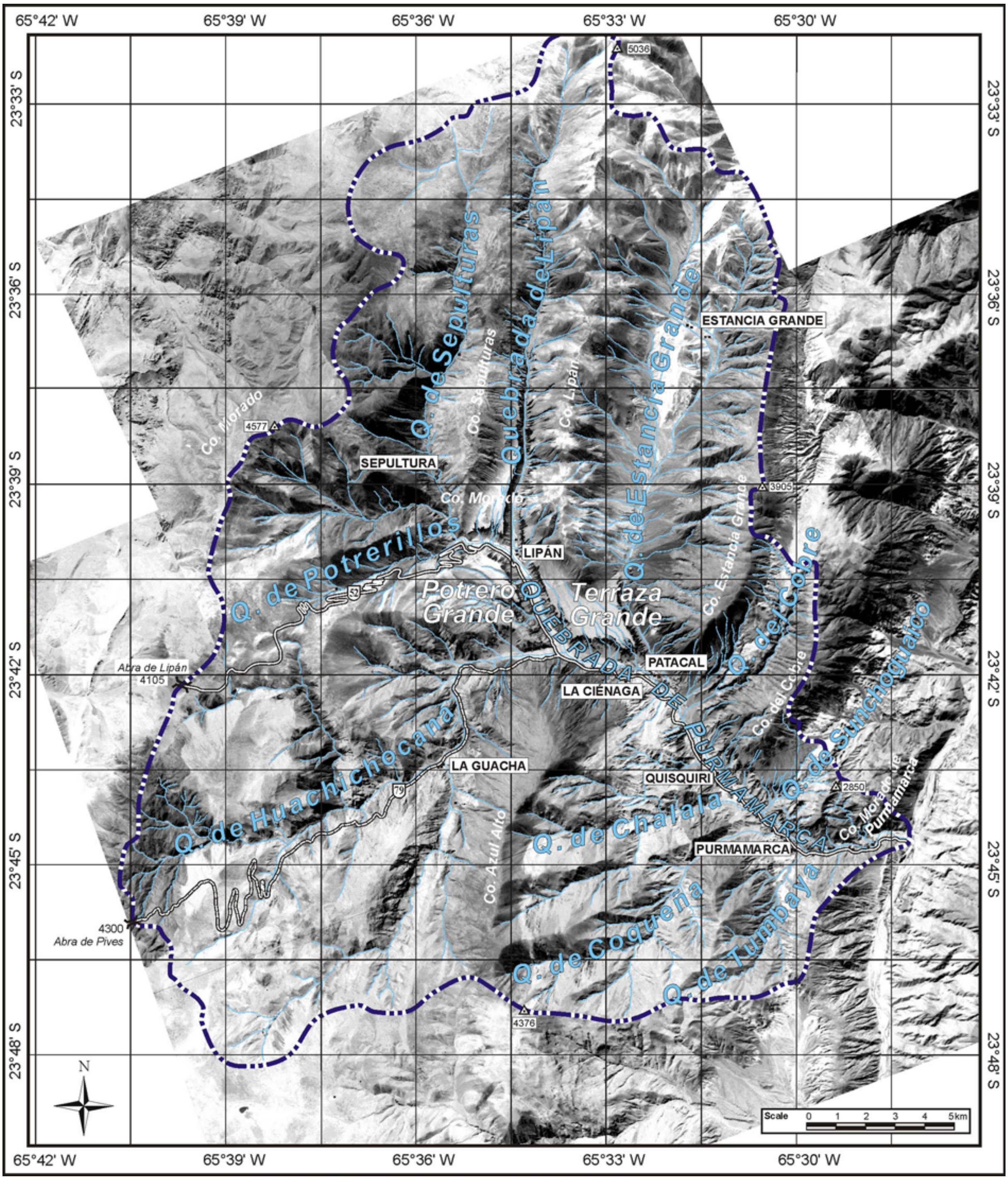


Fig. 3: Mosaic of CORONA satellite imagery of the study area with location names referred to in this study.

2.2. REGIONAL GEOLOGY

2.2.1. GEOLOGICAL EVOLUTION OF THE CORDILLERA ORIENTAL

A short outline of the geological history of the Cordillera Oriental is presented to introduce to the complex regional geology of the study area. The geological history in the Cordillera Oriental begins with the deposition of the Precambrian to Lower Cambrian *Puncoviscana Formation* in a deep-marine environment onto Precambrian crystalline basement (RAMOS ET AL. 1967, AMENGUAL AND ZANETTINI 1974, MON AND SALFITY 1995). While the Puncoviscana Formation underwent folding and metamorphism during the Assynthian orogeny, several intrusions took place throughout the Cordillera Oriental (AMENGUAL AND ZANETTINI 1974). Subsequent uplift followed by erosion and denudation created an extended land surface. On this surface the Cambrian transgression took place, during which the *Mesón Group* was deposited in a near shore shallow-marine environment (RAMOS ET AL. 1967). The Iruyan phase deformed these beds during the Caledonian orogeny, followed again by the formation of a land surface and by the Lower Ordovician transgression, which results in deposition of the offshore-marine *Santa Victoria Group* (TURNER AND MON 1979). Its sedimentation was ended by the Ocloyic phase of the Caledonian Orogeny (TURNER AND MON 1979), which once again was followed by another phase of widespread planation. Whether this phase was interrupted by sedimentation, e.g. during Silurian, Devonian or Carboniferous, is not known for the study area. Unlike in the eastern Cordillera Oriental no evidence has been preserved in its western part (AMENGUAL AND ZANETTINI 1974). The next evidence of deposition comes from the Upper Cretaceous to Lower Tertiary *Salta Group*. Its succession of continental-fluvial, shallow-marine and continental-lacustrine sediments reflects the sedimentary cycle of rift and post-rift processes in the area (SALFITY AND MARQUILLAS 1994). The still continuing Upper Tertiary to Pleistocene *Andean Orogeny* ended deposition and even led to a structural inversion (transformation of normal faults to thrust faults due to compression). Thick alluvial sediments accumulated in the intramontane tectonic basins. Today's structure and morphology is to a large extent controlled by this last and still on-going phase of mountain building (RAMOS ET AL. 1967, TURNER AND MON 1979, MON AND SALFITY 1995).

2.2.2. STRATIGRAPHIC AND LITHOLOGICAL OVERVIEW

The oldest rocks which crop out in the study area are of Upper Precambrian age. These are the low-grade metamorphic rocks of the *Puncoviscana Formation*, consisting of dark greenish-grey to violet quartzitic schists, phyllites and slates, frequently interspersed with quartzitic veins (RAMOS ET AL. 1967, TURNER AND MON 1979). Within the study area the thickness of the formation is more than 1.000 m (RAMOS ET AL. 1967). The *Cambrian Mesón Group,* which overlies the Puncoviscana Formation with a pronounced angular unconformity

(TURNER 1970), is subdivided into the Lizoite, Campanario and Chalhualmayoc Formations (TURNER AND MON 1979).

In the study area a basal conglomerate followed by white and greyish, very hard, fine grained quartzitic sandstones and quartzites are characteristic for the Lizoite Formation. The Campanario Formation mainly consists of light reddish to greyish, fine-grained quartzitic sandstones and quartzites, with thin intercalations of greenish and purple clay- and siltstones. Finally, the whitish fine to medium-grained quartzitic sandstones and quartzites of the Chalhualmayoc Formation are similar to the sandstones of the Lizoite Formation (RAMOS ET AL. 1967, AMENGUAL AND ZANETTINI 1974). RAMOS ET AL. (1967) estimate the thickness of the entire Mesón Group in the study area as being up to 2,100 m, while MON ET AL. (1997) assign only 800 m.

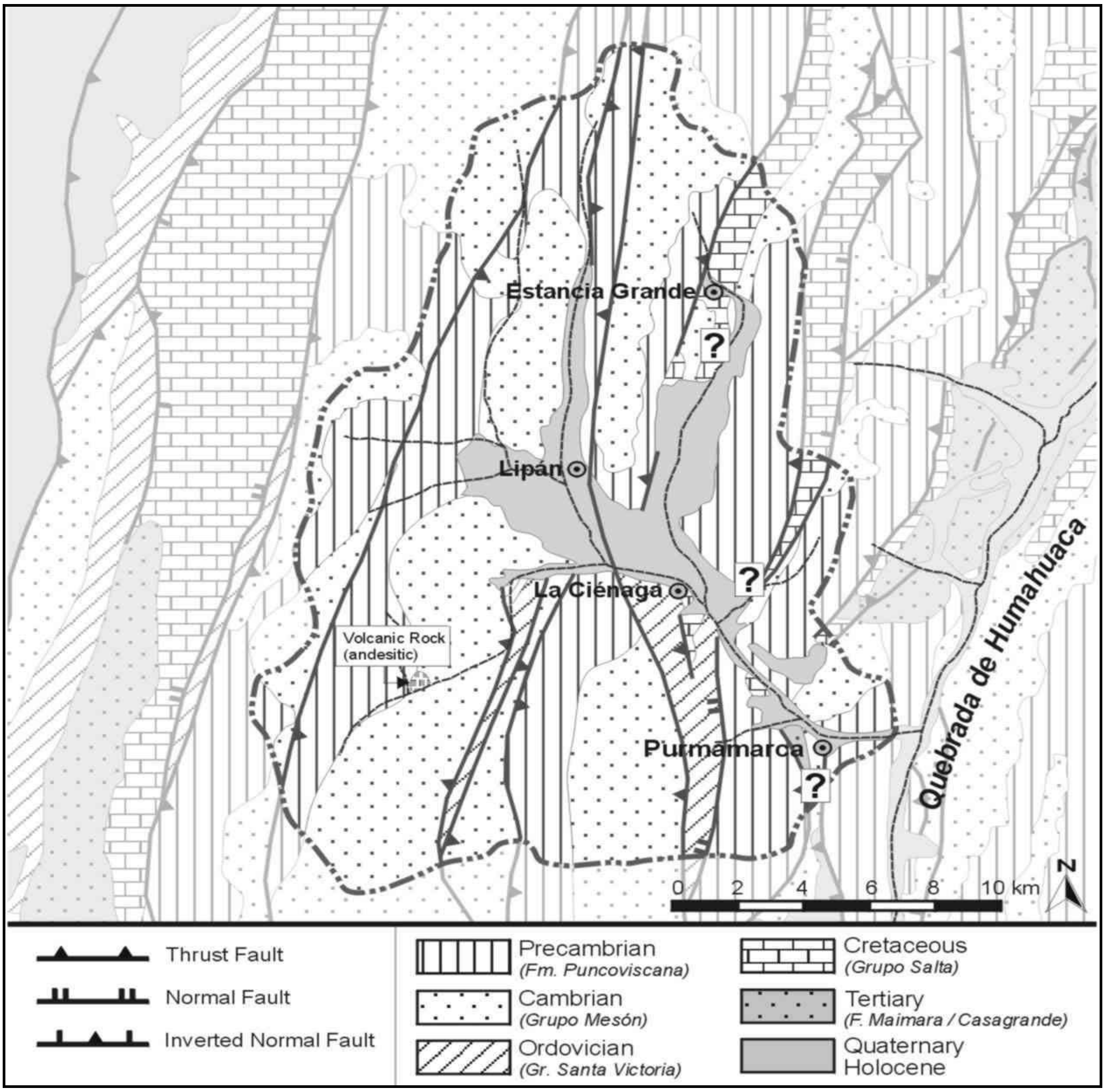


Fig. 4: Geological overview of the study area, "?" denote areas where available geologic maps did not match field observation (map compiled by author, geological data from SEGEMAR 1998 and SEGEMAR-ITGE 1998B).

The Ordovician sediments of the *Santa Victoria Group* are predominantly pelites and greywackes of yellowish to greyish color with few intercalations of sandstones and limestones (TURNER AND MON 1979). They are rich in fossils (RAMOS ET AL. 1967) and overlie the rocks of the Mesón Group in a weak angular unconformity (RAMOS ET AL. 1967, MON ET AL. 1997) The thickness of the Santa Victoria Group for the study area is given as approximately 1,500 m (AMENGUAL AND ZANETTINI 1974).

The Upper Cretaceous to Lower Tertiary *Salta Group* is divided into several subgroups and formations (TURNER AND MON 1979). In the study area only the shallow marine Yacoraite Formation crops out. It rests in transgressive contact above the older formations and consists of light greyish to yellowish limestones interbedded with marls or carbonatic sandstones (RAMOS ET AL. 1967). Outcrops in the study area reach a thickness of 70 – 80 m (RAMOS ET AL. 1967, MON ET AL. 1997?).

In the lower part of the study area outcrops of a reddish carbonatic sandstone interbedded with reddish to brownish clays and interspersed with veinlets of gypsum are (TURNER 1970) have been ascribed to various formations, such as the Upper Cretaceous Pirgua Subgroup (SEGEMAR-ITGE 1998) or Lower Tertiary Santa Barbara Subgroup (SEGEMAR 1998). However, most likely they belong to the Miocene-Pliocene *Chaco Formation* (RAMOS ET AL. 1967, TURNER 1970, AMENGUAL AND ZANETTINI 1974). The Chaco Formation has been called as Casagrande or Rio Grande Formation by some authors (SEGEMAR-ITGE 1998) and Maimará Formation by SALFITY ET AL. (1984). In the study area RAMOS ET AL. (1967) describe the Chaco Formation to crop out in tectonical contact to older formations, but TURNER (1970) confirms an unconformity between the Chaco Formation and older rocks for the Cordillera Oriental. Its thickness is estimated to 600 m in the study area (RAMOS ET AL. 1967).

Of Upper Pliocene to Lower Pleistocene age is the *Uquía Formation*. It consists of very poorly consolidated, fine-grained conglomerates, but also greyish sandstones and greenish to greyish and reddish silts and clays (AMENGUAL AND ZANETTINI 1974, TURNER AND MON 1979). MARSHALL ET AL. (1982) have dated mammal remains from the Uquía Formation to between 2.78 and 1.5 Ma. For the study area RAMOS ET AL. (1967) note a thickness of less than 30 m.

A thick *series of coarse conglomerates* with interbedded lenses of sandy to silty and clayey deposits in the study area is likely to be of Pleistocene age, forming a number of terrace levels (RAMOS ET AL. 1967, AMENGUAL AND ZANETTINI 1974). These are subject to further and more detailed analysis in chapter 4. Some authors (SEGEMAR-ITGE 1998) call them the Purmamarca Formation, but others refer to them simply as "*Pleistocene*" (RAMOS ET AL. 1967) or "*Quaternary alluvial deposits*" (TURNER AND MON 1979), as long as their origin and age have not been confirmed in more detail.

The youngest sediments in the study area are *alluvial and colluvial deposits* similar to the conglomerates mentioned above. To a great extent they are a product of recent reworking processes and fill the valley floors. At many places they rest unconformably above the Pleistocene conglomerates (RAMOS ET AL. 1967).

Finally, RAMOS ET AL. (1967) and TURNER AND MON (1979) mention the outcrop of an isolated *subvolcanic body of andesitic composition* that has intruded the sediments of the Puncoviscana and Lizoite Formations in the upper part of the study area (*Quebrada de Huachichocana*). It is assigned to the Lower Quaternary by correlation to known volcanic activity in the surroundings of the study area, but this can only be considered as a rough first estimate. However, its mineralogical and petrologic properties point to an extrusive nature of the andesite, a fact that could imply that it might have covered a considerable area of a former land surface.

2.2.3. TECTONIC AND STRUCTURAL FRAMEWORK OF THE STUDY AREA

Several superimposed tectonic stages are recognized in the study area, each of them bearing characteristic structural styles of different age (MON AND SALFITY 1995). The oldest distinguishable structures are west-vergent folds and east-dipping faults dating from the intense Paleozoic *Ocloyic Orogeny* that affected Precambrian, Cambrian and Ordovician rocks (MON ET AL. 1997). *Lower Cretaceous rifting* caused subsidence and the opening-up of a series of grabens and hemi-grabens. It generally resulted in N-S and NE-SW trending normal faulting, partly reactivating older structures (MON AND SALFITY 1995, MON ET AL. 1997).

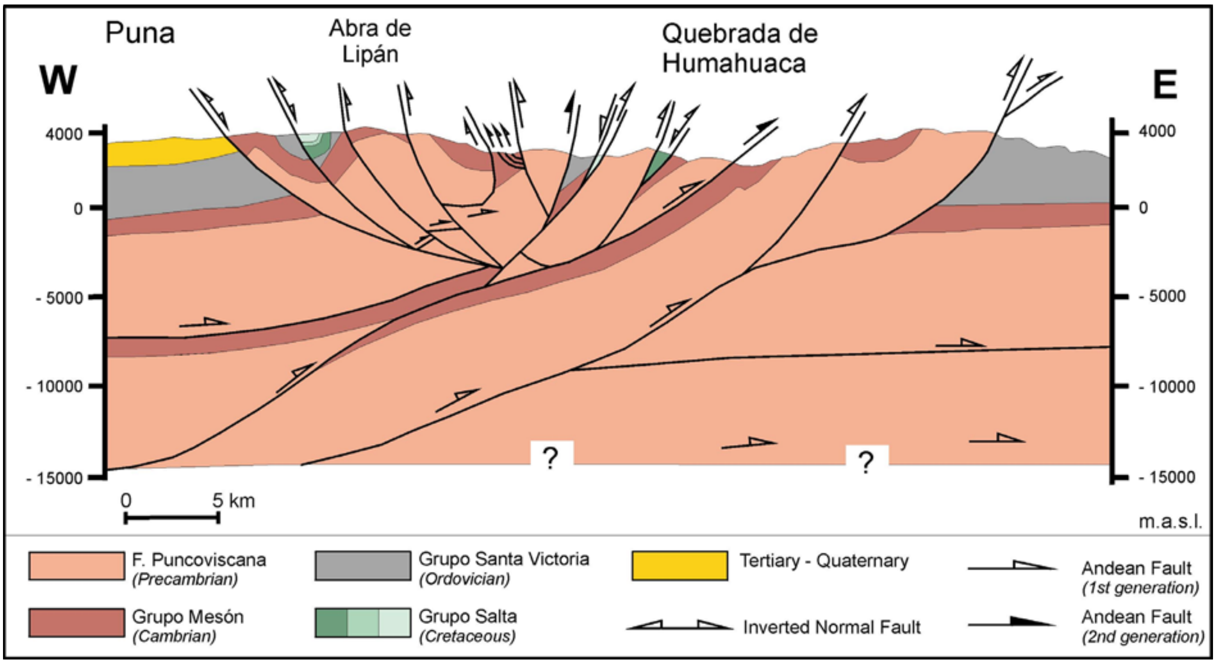


Fig. 5: Structural cross-section of the Cordillera Oriental at Purmamarca, approximately 23° 5' S (modified from SEGEMAR-ITGE 1998A).

Most important for the present structural characteristics of the study area is the *Andean Orogeny*. It commences during the Late Eocene to Lower Oligocene *Incaic phase,* during which the area was uplifted, interrupting sedimentation and causing erosion (SALFITY ET AL. 1996).

Major folding and erosion did not occur until the Late Miocene to Pliocene *Quechua phase* (HARRIS AND MIX 2002, SALFITY ET AL. 1996). Two pulses are distinguished (SALFITY ET AL. 1996), the first of which created a topography pronounced enough to establish the internal drainage of the Puna highlands at 13 – 14 Ma BP (VANDERVOORT ET AL. 1995). Due to subsequent eastward migration of thrusting, uplift was active across the entire width of the Cordillera Oriental by approximately 10 Ma BP, and drainage corresponded roughly to the present pattern (REYNOLDS ET AL. 2000). However, most of the 2,000 meters of uplift in the Cordillera Oriental are reported to have taken place after 10 Ma BP, possibly until about 3 Ma BP (KENNAN 2000). The *Quechua phase* sets the stage for ultimate Andean uplift and directly results from interactions of the Nazcan and South American plates (MON AND SALFITY 1995, MARRETT AND STRECKER 2000).

Finally, the pronounced and compressive Pliocene to Pleistocene *Diaguita phase* generates the eastward folding and faulting of Cretaceous and Cenozoic sediments and reactivates pre-Cretaceous basement, partly inverting Cretaceous rift structures (SALFITY AND MARQUILLAS 1994, MON AND SALFITY 1995). Although already commencing in Mid-Pliocene times, the climax of regional uplift of the Diaguita phase is assumed to have taken place between 1,8 and 1,2 Ma BP, followed by extraordinary high sedimentation rates (HERNANDEZ ET AL. 1996). Even younger regional tectonic events have been estimated as younger than 1 Ma (MARRETT AND STRECKER 2000) for the Cordillera Oriental in NW-Argentina, 600 ka BP for the Sierras Pampeanas as well as the Bolivian Altiplano (STRECKER ET AL. 1989, WIRRMANN AND MOURGUIART 1995) and 250 ka BP for the Sierras Subandinas (HERNANDEZ ET AL. 1996).

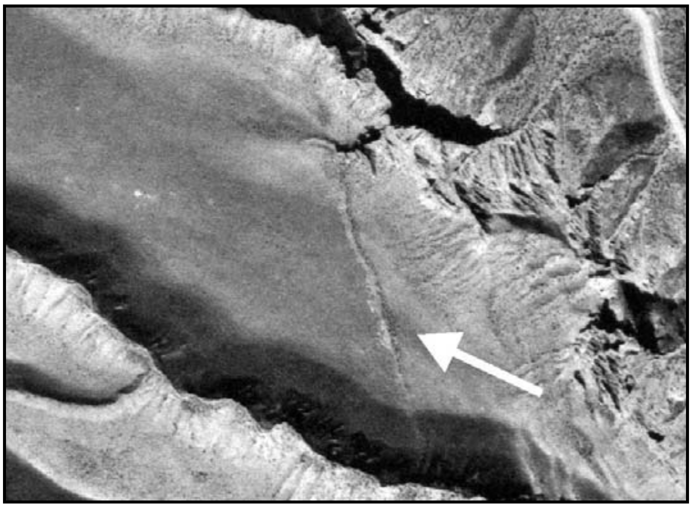


Fig. 6: Possible fault-scarp from recent earthquake activity (GASATACAMA aerial photography, width of image ~600 m)

While changes in fault kinematics have taken place repeatedly during late Cenozoic (SALFITY 1984, MENA 1997, MARRETT ET AL. 1994, MARRETT AND STRECKER 2000), neotectonic activity continues until recent times and manifests itself in minor fault scarps (Fig. 6) and frequent earthquakes (CAHILL ET AL. 1992, HERMANNS ET AL. 2001).

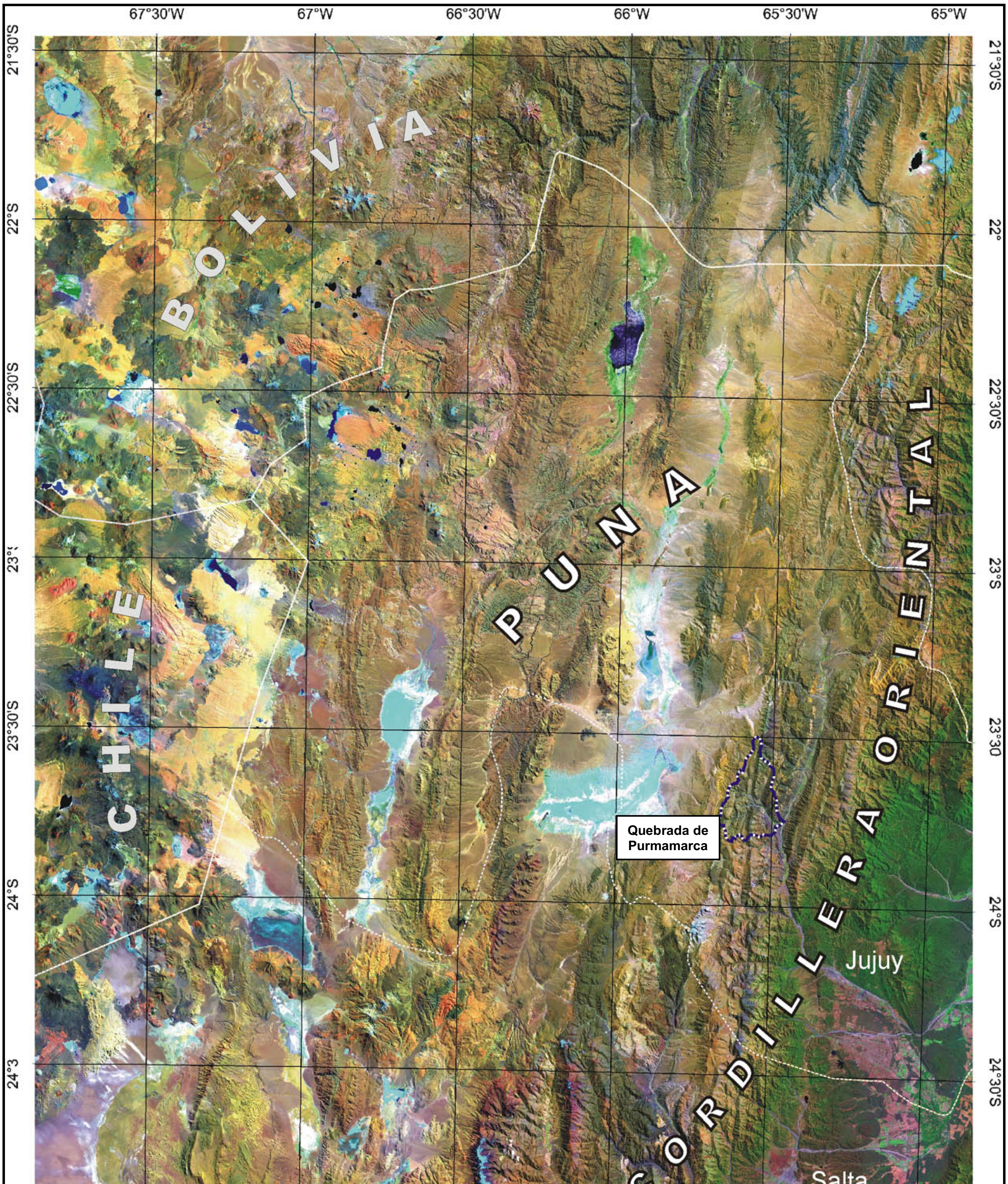


Fig. 7: Mosaic of Landsat TM 5 imagery of NW-Argentina and the study area.

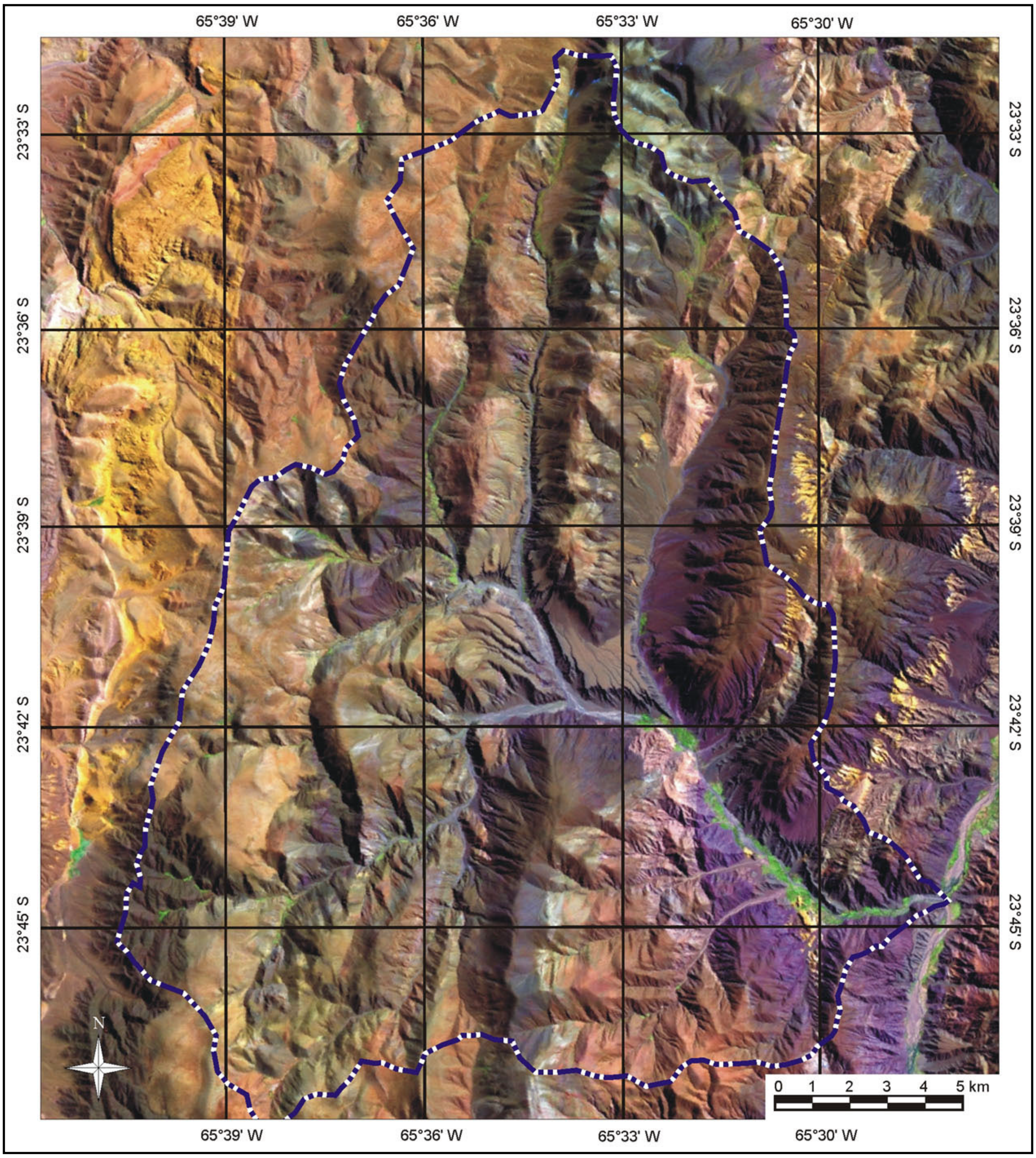


Fig. 8: Landsat 5 TM image of the study area, band combination 7-4-1. Geologic and lithologic differences show up particularly well due to their selective radiation characteristics. In this respect, the Precambrian Puncoviscana Formation (violet) and Cambrian Grupo Mesón (brownish) constitute an important part of most N-S orientated fault-blocks. Structurally, three main directions are clearly distinguishable by the orientation of the valleys: N-S, NW-SE and ENE-WSW. In the topographically higher parts in the west of the basin, colors are not as manifold. Slopes and ridges are covered by a thicker mantle of slope debris. Agriculture and vegetated areas shine in a light green and are concentrated to small parts of the valley floor. Only in the higher parts of the catchment they can be found on fans and terraces, which is certainly an expression of local climatic variations.

2.3. MORPHOLOGICAL FRAMEWORK

The Quebrada de Purmamarca is a drainage basin to the west of the Quebrada de Humahuaca in the western part of the Cordillera Oriental, tributary to the Rio Grande basin, which is one the most important drainage basins in NW-Argentina. The overall topography of NW-Argentina is basically controlled by the transition between the orogenic body of the Andes to its foreland (Fig. 7).

From a morphostructural point of view the *Cordillera Oriental* is a fold and thrust belt of mainly Paleozoic rocks with a typical piggy-back (intramontane basins in active thrust belts) structural style (MON AND SALFITY 1995). Its ranges are usually associated with high-angle thrust faults and reach altitudes of well over 5,000 m.a.s.l.. Morphologically the ranges alternate with alluvial valleys. Just like their most important example, the *Quebrada de Humahuaca*, most valleys seem to have developed and incised along lines of structural weakness (IGARZABAL 1991).

An enormous thrust with as much as 15 km of displacement in the eastern direction marks the eastern border of the Cordillera Oriental (MON AND SALFITY 1995). The transition to the Andean foreland *(the Chaco lowlands)* is constituted by the *Sierras Subandinas*, a fold belt with ranges averaging 1,000 m.a.s.l. (MON AND SALFITY 1995). Morphologically the Sierras Subandinas consist of fold ranges. Anticlines commonly correspond to ranges and valleys have formed in the synclines. Due to its position within the atmospheric circulation it receives high amounts of precipitation leading to a thick vegetational cover (IGARZABAL 1991). A dense drainage net has developed since the initial uplift. To the west, the Cordillera Oriental is bounded by the *Puna plateau* with average elevations of 4,000 m.a.s.l.. Structurally this plateau consists of a Cenozoic volcanic arc and eastward-thrusted blocks of Paleozoic rocks. It is characterized by a basin and range morphology, internal drainage and important salt pans (IGARZABAL 1991, MON AND SALFITY 1995).

Therefore an enormous relief with high elevation differences characterizes the Quebrada de Purmamarca (Fig. 8). While its confluence into the Quebrada de Humahuaca is located at approximately 2,200 m.a.s.l., its highest peak reaches 5,036 m.a.s.l. within a distance of only 25 km. The major valley of the Quebrada de Purmamarca, branches out into several lateral valleys. The most important of these are the Quebrada de Huachichocana, Quebrada de Estancia Grande, Quebrada de Lipán, Quebrada de Sepulturas and Quebrada de Potrerillos. The valleys are divided by mountain chains of varying elevations between 2,800 m.a.s.l. and 5,000 m.a.s.l., while the highest chains are towards the west, where the Cordillera Oriental transits into the Puna plateau. With an orientation roughly N-S to NE-SW most mountain chains reveal strong structural controls. Remarkable terrace systems characterize the deep valleys. Despite its pronounced relief and enormous range of elevation, the overall geomorphological appearance of the mountain chains is relatively smooth.

2.4. REGIONAL CLIMATE

2.4.1. CLIMATIC CHARACTERISTICS OF NW-ARGENTINA

The regional climate of northwestern Argentina is controlled by two main factors. Within the global atmospheric circulation system NW-Argentina is positioned at a transitional situation between tropical and subtropical, that is to say continental and Pacific influences. In addition, a range of regional climatic phenomena is caused by the large altitudinal differences (300 – 6,880 m.a.s.l.). These generate a strong vertical climatic gradient and are also responsible for the altitudinal zonation, which is particularly evident in vegetational patterns (PROHASKA 1976, WEISCHET 1988).

Four dynamic components of the global atmospheric circulation system are climatically relevant for NW-Argentina. These are the subtropical high-pressure cell of the southeasterly Pacific Ocean, the high-pressure cell of the southern Atlantic Ocean, the ITCZ of low pressure and the southernhemispheric zone of the westerlies (ENDLICHER 1995, WEISCHET 1996).

Despite the relatively short distance to the Pacific Ocean and the Pacific high-pressure system most of its influence on NW-Argentina is literally blocked by the Andes. Thus the dominant air masses reaching the area are of very humid-tropical-continental or Atlantic origin (WEISCHET 1988). Only on a few occasions the zonda, a westerly wind, brings warm and dry air masses to NW-Argentina (ENDLICHER 1995, WEISCHET 1996).

During austral winter the climatic situation is frequently characterized by high pressure that forms a bridge between the high-pressure cells of the southern Atlantic and that of the Pacific Ocean. Little precipitation and cloud cover, high insolation but relatively cool temperatures are characteristic. In addition, infrequent low-pressure cells can cause cold fronts to determine the regional weather, when cyclones make their way from the southern zone of the westerlies east of the Andes up to NW-Argentina (ENDLICHER 1995).

Throughout the entire year tropical maritime air masses can bring humidity from south-easterly directions, being transported alongside the Atlantic high pressure cell and causing light rainfall of some duration (ENDLICHER 1995).

During austral summer NW-Argentina lies under the influence of a strong continental low-pressure cell. This is caused by strong insolation and convection and has therefore been considered to be a southern extension of the ITCZ. Nevertheless its formation can also be attributed to pronounced heating above the elevated Puna plateau as well as the effects of wind shadow of the Andes (ENDLICHER 1995). Whatever its origin, this low pressure cell takes in warm and moist tropical continental air masses from the N and NE which supply the region with humidity and cause efficient rainfall in summer, accounting for 70 – 90 % of the total annual precipitation (WEISCHET 1991, MARTYN 1992).

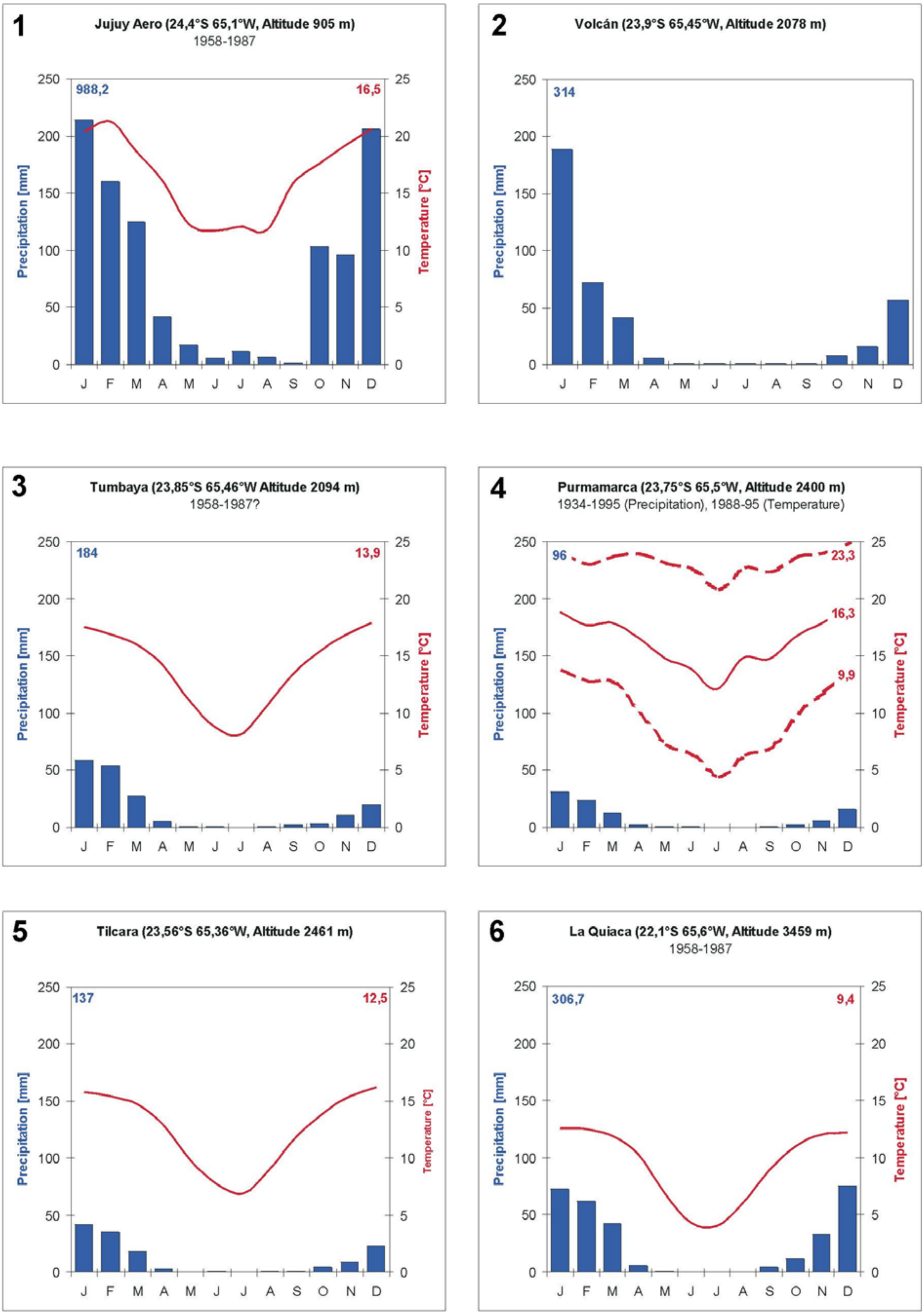


Fig. 9: Climatic diagrams of selected stations in the province of Jujuy (mean monthly precipitation and temperature, mean maximum and mean minimum monthly temperature for Purmamarca), numbered from S to N and from lowest to highest elevation. No.2-6 located in the Quebrada de Humahuaca. Note low annual total precipitation in central part of the quebrada (No.2-5) and pronounced temporal distribution of precipitation (Data from NCDC, CHOROLQUE 1998 and SOLER 2002).

Due to great differences in elevation and different intensities of insolation effective local and regional wind systems have a strong influence on the local climate and interfere with regional climate patterns. Commonly cold downslope air flow takes place through the night and morning, before turning to the uphill directions in the afternoon and evening (ENDLICHER 1995). On a seasonal basis a regional circulation system causes air mass exchange between the Puna Plateau and the Chaco lowland, with directions towards the Puna in winter and towards the Chaco in austral summer. It is caused by seasonally varying amounts of insolation (PROHASKA 1976, WEISCHET 1996).

2.4.2. PRECIPITATION

From a global point of view the climate of NW-Argentina has a subtropical to tropical climate of summer rain (PROHASKA 1976, MARTYN 1992). As evident from the climatic diagrams in FIG. 6 precipitation in NW-Argentina is highly seasonal; rain falls mainly in summer, - between November and February. Following KÖPPEN'S classification this would be a Cwb (*Jujuy*) and BSk (*La Quiaca*) to BWk (*Purmamarca*) climate, the latter being semi-arid to arid (LAUER 1995). Usually most of the precipitation falls as heavy rain- or thunderstorms of relatively short duration (WEISCHET 1988). In general the annual quantity of precipitation decreases from east to west, due to the increasing influence of the South Pacific subtropical high (MARTYN 1992), but at a regional level the precipitation pattern is determined by regional topography and elevation. Therefore the amount of rain generally decreases with increasing altitude (Fig. 9: *Volcán, Tumbaya, Tilcara*).

As yet there are only unofficial climatic data from private recordings for the Quebrada de Purmamarca (1988-1995 for temperature, 1934-1995 for precipitation), taken at the village of Purmamarca (personal communication R. SOLER). Whether they are representative for the entire Quebrada or whether local topographic differences generate a much more complicated climatic pattern cannot be decided without additional data. It seems safe to assume that the temporal distribution of rainfall is generally similar to the regional pattern, but spatial distribution varies due to local topographical influences. Particularly the deeper quebradas have been observed to receive less precipitation than the surrounding mountain slopes, a phenomenon attributed to the compensation of thermic differences (WEISCHET 1996).

Still it should be mentioned that even though the major amount of rain falls during strong summer storms, light rain of longer duration is not uncommon and presumably of great importance for soil formation and geomorphic processes (OWN OBSERVATION). With an average of 96 mm per annum, precipitation is even lower than that of stations outside the Quebrada de Purmamarca. This may be due to the relatively narrow and steep entrance to the Quebrada, which constitutes an obstacle for air masses coming up the Quebrada de Humahuaca.

2.4.3. TEMPERATURE

Temperatures in NW-Argentina are strongly controlled by elevation as well as exposure, the nature of the ground surface and the amount of insolation reaching it. Therefore in topographically lower stations maximum and minimum annual temperatures show high thermal amplitude throughout the year (Fig. 9 *Jujuy*). In contrast, for topographically higher stations strong daily variations are more characteristic (Fig. 9 *La Quiaca*).
For NW-Argentina PROHASKA (1976) notes the average height of the 0° C annual isotherm to be between 4.500 and 5.400 m.a.s.l.. Intra-annual average temperatures at Purmamarca show an amplitude of approximately 20°C. Daily variations are of considerable importance, particularly in winter when insolation is greatest due to low cloud cover. Average monthly minimum temperatures may drop to below 5°C even in the lower part of the Quebrada, which implies average winter temperatures at higher elevations well below freezing point (Fig. 9). In the lower part of the Quebrada the period of days with possible frost averages 165 days between November and March. With increasing elevation this value increases to over 250 days (CHOROLQUE 1998).

2.4.4. ENSO INFLUENCE

The influence of the ENSO (El Niño Southern Oscillation) phenomenon on the regional climate which is known from many parts of South America (DIAZ AND KILADIS 1992), can also be observed in NW-Argentina (PRIETO ET AL. 1998, TRAUTH AND STRECKER 1999), although the impact is not as dramatic as elsewhere. The ENSO impact is particularly well reflected by precipitation anomalies.
For NW-Argentina's intra-Andean mountainous regions TRAUTH ET AL. (2000) report drier weather during El Niño and wetter conditions than normal during La Niña events. The ENSO impact for the lowland regions east of the Andes is just the opposite, with wetter conditions during El Niño and drier ones during La Niña events. A second factor controlling the intensity and amount of summer rains seems to be the decadal variation of interhemispheric sea surface temperatures in the Atlantic (TAD). A minimum TAD amplitude increases the amount of rainfall in virtually all of NW-Argentina, locally by even as much as 200% (TRAUTH ET AL. 2000)

2.5. VEGETATION OF THE STUDY AREA

The vegetation cover does not only reflect the climatic characteristics of a region but also exerts a great influence on the morphodynamic processes of a given area. A short outline of the vegetation cover in the Quebrada de Purmamarca reveals the altitudinal zonation of climate and emphasizes the semi-arid to arid character of NW-Argentina in general. RUTHSATZ (1977) and WERNER (1978) divide regional vegetation into three main communities

belonging to different altitudinal zones. These communities stand for a general picture of regional vegetation rather than presenting a complete record of species.
In the highest part of the study area, up to 4,150 m.a.s.l., vegetation is dominated by the high-Andean *Festuca* bunchgrass community mainly consisting of *Festuca othophylla*. It is characterized by a very sparse cover of around 10 %. Locally small depressions or favorable exposition modify the bunchgrass community towards a higher percentage of ground cover, with the inclusion of dwarf shrubs (e.g. *Parastrephia quadrangularis, Nassauvia axillaris, Baccharis incarum, Azorella compacta*). The transition to the next lower vegetational zone is marked by an increasing number of shrubs (*Oreocereus trollii, Tetraglochin cristatum*), announcing the community of Andean shrub plants, particularly dominated by *Baccharis boliviensis* (LUPO 1998) and *Verbena asparagoides*. These communities populate altitudes between 4,150 and 3,400 m.a.s.l., but can be subdivided into an edaphically more humid part above 3,700 m.a.s.l. and a drier part below 3,700 m.a.s.l., where there is a significant decrease in the number of herbal plants. On north exposed slopes cacti like *Trichocereus pasacana* are also common. The lowest vegetational zone in the study area below 3,400 m.a.s.l. is populated by *Gochnatia glutinosa* -, and *Abromeitiella lorentziana* communities. The *Abromeitiella* communities typically populate north-exposed slopes and are characterized by cushion and carpet plants as well as cacti, while *Gochnatia* communities, in particular interspersed with *Ephedra breana*, are found on south-exposed slopes.
Despite the apparent relation between altitudinal zones, climate and vegetation, WERNER (1978) points out azonal modifications of the vegetational assemblage due to local influences. In this context the habitats of initial and degradational vegetation communities have to be mentioned. They cover relatively large parts of the study area and are subject to strong erosional processes, which are likely to have been triggered and amplified by anthropogenic influences (CHOROLQUE 1998, LUPO 1998).

2.6. SOILS OF THE STUDY AREA

Despite the scarcity of pedogenic research in NW-Argentina, some general observations show the close link of soil type to climate and vegetational cover and reflect the semi-arid to arid character of the study area. Due to the deficit of water and the loose vegetational cover, regional soil formation is usually characterized by very slow pedogenic processes.

If the climatic conditions allow, mineralization and humification of plant material takes place and forms initial A-Horizons (RUTHSATZ 1977). The resulting soil types may be classified as *leptosols* and *regosols*. During austral summer, interflow and descending water percolation within the upper few centimeters to decimeters will transport soluble materials such as chlorides, sulphates and particularly carbonates (WERNER 1971). Therefore several authors have reported soils with characteristic horizons of carbonate accumulation (*Calcisols, Calcretes*) from NW-Argentina (WERNER 1970, WERNER 1971, KRISL 1999). Finally, *cambisols* with typical reddish to brownish, clay rich, carbonate free B-horizon have been observed in parts of NW-Argentina (WERNER 1971, KRISL 1999). ZIPPRICH (1998) even reports *luvisols* for the Sierra de Santa Victoria. Whether these soils actually form under present climatic conditions or represent relicts of a much wetter climatic period is a key question for paleoenvironmental interpretation and the establishment of a regional frame of landscape history (EBERLE 2000). In addition, the soil types in the study area and their association present valuable information for the interpretation of terraces and alluvial fans (chapter 4).

2.7. HISTORICAL OVERVIEW

The name *Purmamarca* is derived from the indigenous language Aymara with *"PURMA"* meaning *"unseeded field"*, *"desert"* and *"MARCA"* meaning *"village".* This way Purmamarca could be translated as *"village of virgin land"* or *"village in the desert"* (CHOROLQUE 1998). The earliest evidence of mankind in the study area has been found in a cave in the Quebrada de Huachichocana. It dates back to as early as 10,000 – 8,000 yr. BP and has been associated with the archeological *preceramic period* (10,800 - ~3,000 yr. BP), a period of hunting and collecting, nomadism and subsequent experimental agricultural activities, initial domestication of animals and trade to regions like the Pacific coast and the Chaco (J.A. KULEMEYER 1998). During the early *formative period* of the Agroalfarero cultures, peoples commenced to form permanent settlement on the base of agriculture and stockbreeding (J.A. KULEMEYER AT AL. 1999, J.J. KULEMEYER 1998). Within the study area a site of more than 200 ha of housing and agricultural ruins has been identified near Estancia Grande (CHOROLQUE 1998). After ongoing political expansion and cultural development in NW Argentina, the region became part of the Incaic empire around 1470 AD, a period archeologically known as the *Incaic period*. Due to its transitional location between the Quebrada de Humahuaca and the Puna and its connection to the Incaic road network, this must have meant a pronounced increase in traffic and importance for the Quebrada de Purmamarca (CHOROLQUE 1998). During this time various aboriginal groups inhabited the region, making their living with irrigation agriculture. The period of initial *Spanish-indigenous contact* (1535 AD – 1600 AD) in NW-Argentina began with the arrival of the expedition of Diego del Amagro, which is thought to have penetrated the Quebrada de Purmamarca on its way down from the Puna (CHOROLQUE 1998, J.J. KULEMEYER 1998). In 1594 AD the village of Purmamarca was founded at the site of a Precolombian settlement (CHOROLQUE 1998). The subsequent period of *Spanish colonization* (1600 AD – 1810 AD) was characterized by enormous social and political changes in all of Latin America, as an agricultural system based on Spanish large estates and an indigenous working force was established (J.J. KULEMEYER 1998). Finally, the still ongoing period since Argentinean *independence* (1810 AD – today) has caused economical and agricultural decline in NW-Argentina, which is the reason for the evident rural exodus from the region (J.J. KULEMEYER 1998).

2.8. POPULATION, INFRASTRUCTURE, AND ECONOMY

The municipality of Purmamarca is part of the department (~county) of Tumbaya in Jujuy, the northernmost province of Argentina (STN 1995). For 1997 its population has been estimated at 1,966 inhabitants. Almost 50 % of these live in Purmamarca, the largest of six settlements with more than 100 inhabitants (CHOROLQUE 1998). The smaller settlements usually consist of several adobe houses scattered along the valley sides and do not have any infrastructure. At some places, they are connected to the road network only by paths. Nevertheless, National Road *Nr. 52* (via Abra de Lipán) and Provincial Road *Nr. 79* (via Quebrada de Huachichocana and Abra de Pives) run through the municipality (IGM 1998). Both are dirt roads, but the latter is now in a very bad condition. The national route is passable throughout the entire year and heavy traffic has intensified to up to 2,000 trucks passing per month (CHOROLQUE 1998). Practically, it constitutes one of two existing passes for heavy traffic connecting Northern Argentina and Paraguay with Chile and is therefore of great importance not only for the province of Jujuy, but for the entire region.

Due to its natural beauty (*"Cerro de los Siete Colores"*) and tranquility, the village of Purmamarca has become a popular place for tourists to stop by. So far tourism has not brought important economical improvements to the area, as it is very transitory and little organized. Nevertheless, Purmamarca is famous for its market and handicrafts. A more important source of income for the entire Quebrada de Purmamarca is the agricultural production of vegetables, legumes, cereals as well as fruits. In most cases production remains on a subsistence level and only small amounts of agricultural goods are exported from the area. Finally, livestock, particularly goats, sheep and, to a lesser extent, cattle graze on the extended pastures along the slopes and terraces of the area. They are subject to a transhumance system and constitute a severe danger of overgrazing (CHOROLQUE 1998).

3. METHODICAL APPROACH

The definition of the study area – the Quebrada de Purmamarca - has been realized following geomorphological criteria: even though a watershed or drainage basin is a hydrological unit, the availability of water is always a factor of geomorphological importance, too. Drainage divides have been mapped from remote sensing data. The Quebrada de Purmamarca is subdivided into several smaller valleys or canyons (Spanish: *quebradas*). For further explanation of names and locations see Figure 3.
As mentioned above, the goal of this study is the reconstruction of a landscape evolution with particular attention to geomorphology. Several geographic methods as well as a variety of data from different geoscientific disciplines find entrance to the study. While this study bases its main conclusions on sedimentological and geomorphological data, landscape evolutionary studies in general can very well be referred to as multidisciplinary.

A first introduction to geological and geomorphological problems of the study area dates back to July 1999. Within the scope of an internship at the National University of Salta, Argentina (UNSa) analyses of air photography as well as a four-day field survey brought up a variety of questions. Particularly due to the lack of geomorphological investigations in the entire region, the question for landscape evolution in the Quebrada de Purmamarca became the subject of this study.
Aside from preliminary literature research, several types of remote sensing data have been acquired and visually interpreted in order to narrow down the possible starting points for field work. In this context, LANDSAT TM 5 data has been particularly useful to gain a geological and geomorphological overview (SIEFKER 2001). More detailed geomorphological information has been derived from CORONA satellite imagery (USGS/EDC 1995). This panchromatic imagery from a former U.S. military satellite active between 1959 and 1972 has been opened to the public in 1995 and is available on a low-cost base. Particularly its high spatial resolution of nine feet (~2,75 meters) makes it extraordinary useful for orientation, mapping and image interpretation. Additional information has been extracted from *topographic and geological maps*. Topographic maps for the study area are only available at the scale of 1:250.000 (IGM 1986), while geological maps exist at a scale of up to 1:100.000 (SEGEMAR 1998, SEGEMAR-ITGE 1998).

However, the central part of this study have been three months of field work carried out between March and May 2001. Geomorphological inquiry and sedimentological and pedological analysis have been the main focus of field work. In this respect GPS technology had been applied for orientation and a precise measurement of topography (for GPS data see table 6, appendix). Elevation measurements have additionally been adjusted with a barometric altimeter. All observations have been photographically documented. Sedimentological-stratigraphical profiles have been described in detail at 15 locations and a total number of 31 samples have been taken for granulometric analysis. In addition, paleoflow measurements have been taken at five profiles. At 12 locations pedological profiles have been described and a total number of 19 samples have been taken for granulometric and $CaCO_3$-analysis. In addition, several hand samples of soil components have been taken for micromorphological analysis.

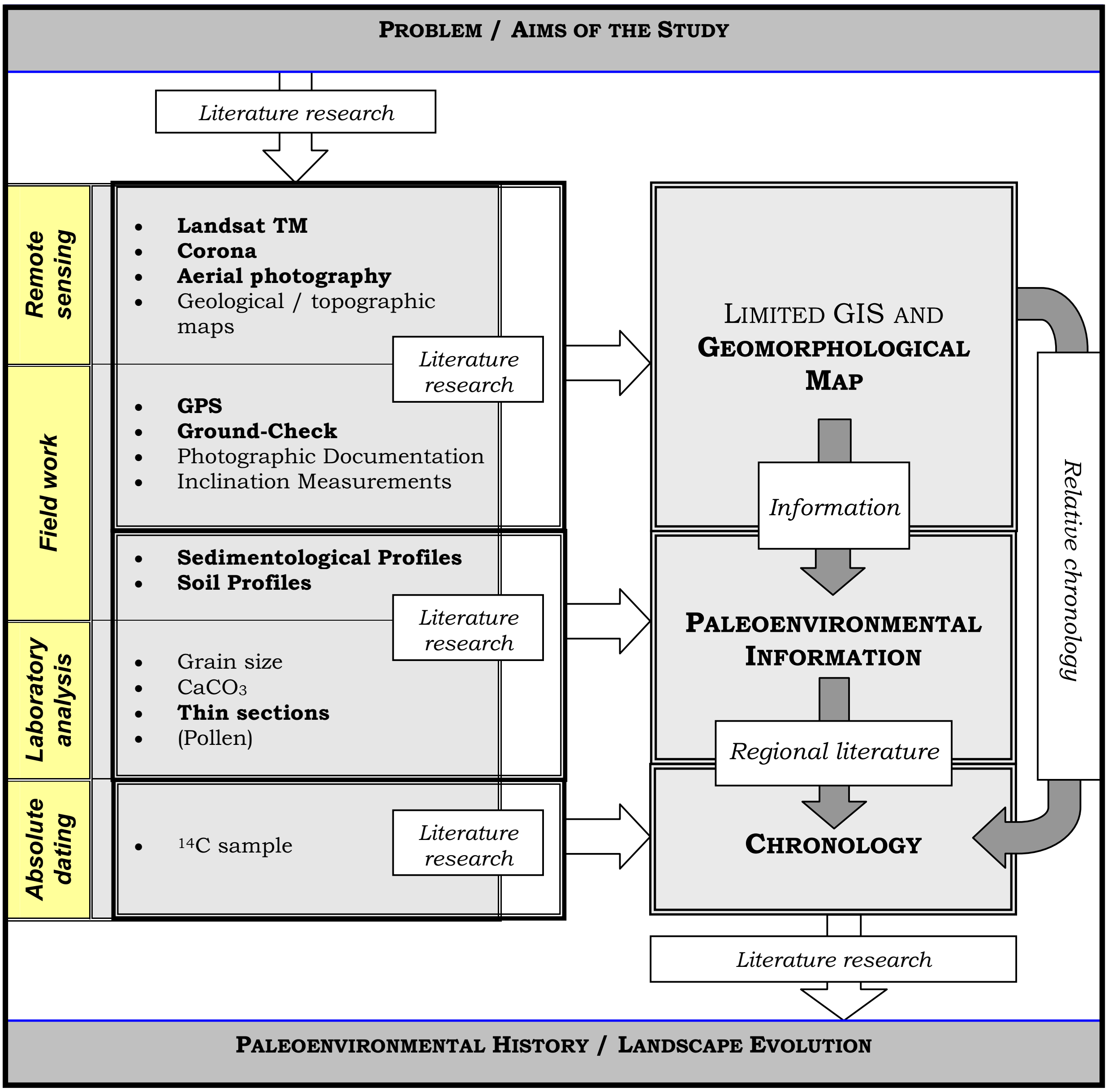


Fig. 10: Methodical framework of this study.

In July 2001 following the field work, most sedimentological analysis of granulometry and $CaCO_3$-content was carried out in the geomorphological laboratory of the University of Würzburg. All sedimentological analysis followed basic geomorphological methods (e.g. Leser 1977). Preparation as well as analysis of seven samples for micromorphological investigation has been done in the mineralogical institute.

In order to evaluate all geomorphological observations from the study area, several methods have been applied. Aerial photography as well as stereopairs of CORONA imagery have been practical for stereoscopic interpretation. Due to its extraordinary well spatial resolution, aerial photography from GASATACAMA resulted to be particularly useful and allowed in some cases the recognition of geomorphological features on a scale of few meters. GPS data has been processed and organized using ARCVIEW 3.2 software. Satellite and map data has been digitized and georeferenced using ENVI 3.5 software. All field observation have been included into the resulting limited GIS (Geographic Information System).

In addition, a geomorphological map at an approximate scale of 1:85,000 has been compiled on the base of geomorphological observations. It mainly integrates the results from field work and image interpretation from remote sensing data and can be considered a basic geomorphological GIS, a fact which is particularly evident from the arrangement of different types of information in different layers. A georeferenced mosaic of two scenes of CORONA imagery has been used as the base map. The task of combining various types of geomorphological information, assigning them to different layers and graphically designing them has been implemented using COREL DRAW 9.0 software.

INFORMATION LAYER	GEOMORPHOLOGICAL FEATURES
MORPHOGRAPHIC INFORMATION	Ridges Scarps Fan (alluvial) Plain (erosional, depositional)
MORPHOGENETIC INFORMATION	Denudational Denudational-Gravitational Fluvial-Denudational Fluvial-Alluvial Fluvial
PROCESS-MORPHOLOGICAL INFORMATION	Erosion (lateral erosion, incision, gullying) Aggradation Deflation
HYDROGRAPHIC INFORMATION	Drainage lines (perennial, periodic)
OTHER INFORMATION	Settlements, roads, agriculture, peaks, passes

Table 1: Information layers of the presented geomorphological map and relevant geomorphological features.

According to existing methods for the production of geomorphological maps (LESER AND STÄBLEIN 1975, BARSCH AND LIEDTKE 1980, STÄBLEIN 1980), the presented map consists of five different layers (Table 1). The most relevant layer for interpretation regarding the landscape evolution is certainly the layer of morphogenetic information. Besides the illustration of results from regional geomorphological research, geomorphological maps form the base for a variety of applications in many disciplines (WALZ 2001). Therefore the geomorphological map presented can be considered an essential product of this study.

One main focus of the interpretation of the observations and results has been the question as to what extent landforms, sediments and soils and the processes of their formation reveal paleoenvironmental information. Besides being based on general principles and knowledge, the interpretation of the geomorphological, sedimentological and pedological results has to some extent required a comparison with existing literature.

Equally important has been the establishment of a chronological frame for the results obtained. In a first step a relative chronology has been set up according to principles of geomorphological and stratigraphical principles. The assignment of absolute dates has been possible only considering certain restrictions. At several places of the study area material suitable for absolute age dating has been discovered. From an outcrop at Potrero Grande a greyish and consolidated tuff has been extracted and a whitish layer of ash deposit has been found in the Quebrada de Sepulturas. Unfortunately none of these volcanic deposits has proved suitable for absolute age dating due to their poor preservation and advanced state of weathering (personal communication ANDREAS RICHTER, *University of Potsdam*). However, some results may be obtained if the superficial weathering of these deposits is considered during future sampling. More detailed tephrostratigraphic correlation (e.g. BOLLI ET AL. 1998) might also bring some results.

In addition, at several localities gastropods and plant remains have been extracted from sedimentological profiles. Theoretically, fossil material of certain carbon content is suitable for absolute age dating. Nevertheless, from three samples only a single date has been obtained. The other samples have been found to contain too little organic carbon for precise age dating (personal communication ANDREAS RICHTER, University of Potsdam). The ^{14}C-date obtained from Prof. Dr. P. M. Grootes at the "*Leibniz Labor für Altersbestimmung und Isotopenforschung*" at the University of Kiel (appendix) has been declared to be 49,550±1700 years old.

The final part of the study has been concerned with the thematic and graphic representation of the results as well as with their interpretation regarding the landscape history of the Quebrada de Purmamarca. In this respect several cross-references have been included in the discussion to confirm the results and their interpretation within a regional framework.

4. FIELD EVIDENCE FROM THE QUEBRADA DE PURMAMARCA

In this chapter, the results of geomorphological observations from field work as well as additional implications from analysis of remote sensing data will be presented. Where possible they are interpreted and discussed regarding paleoenvironmental questions. The geomorphological map (*see end of this volume*) serves to summarize these results and put them in context to each other.

4.1. PALEOPLANATION SURFACES

In the study area, several mountain chains exhibit an extremely flat topography. These mountain tops are of relatively narrow but long extent, usually a few hundred meters broad and up to several kilometers in length. Elevation does not vary more than a few tens of meters over this distance. The observed flat mountain tops all show very different altitudes, ranging from 2,800 m.a.s.l. in the eastern study area to 4,600 m.a.s.l. in the western study area. This means that they are found at increasingly higher altitudes the further west they are located. Despite their varying altitude, all of the observed flat mountain tops seem to follow a NNE-SSW orientation. Considering the structural characteristics of the entire region (2.2.3.), the orientation of most mountain chains corresponds to high angle, NE-SW and N-S trending thrust faults. Table 2 summarizes the identified mountain chains, their orientation and their approximate altitude.

LOCATION WITHIN THE STUDY AREA	ORIENTATION	ALTITUDE [m.a.s.l.]
Qbd. de Tumbaya (East)	NNE-SSW	2,800
Qbd. del Cobre (Southeast)	NNE-SSW	3,400
Qbd. de Estancia Grande (East)	NNE-SSW	3,850
Qbd. de Chalala / Coqueña (West)	NNE-SSW	3,900 – 4,300
Qbd. de Sepulturas (West)	NNE-SSW	4,600

Table 2: Summary of supposed remnants of paleoplanation surfaces on flat-topography mountain chains (from E to W).

The morphology of these mountain chains suggests that they can be considered to be remnants of paleoplanation surfaces. In any case, climatic and geomorphological conditions and processes involved in the formation of a planation surface differ significantly from the processes observed today. The flat topography implies a dominance of significant denudational processes for the formation of the planation surfaces. Other than the morphological appearance of the paleoplanation surfaces, no saprolite or other evidence for processes involved in their formation has been observed in the study area. Therefore it is difficult to assign certain denudational processes to the formation of the surfaces.
However, the paleosurface has formed in environments much lower than the present topography. Consequently, the great amount of uplift to the present elevations can be attributed the structural and tectonic characteristics of the study area (OLLIER AND PAIN 2000, BURBANK AND ANDERSON 2001). Most thrust faults show the same orientation as the

remnants of the paleosurfaces. In the western study area, close to the Puna plateau, elevations of the paleosurfaces are highest. Here, thrusting must have been most intense and has therefore probably continued for longer time.

Fig. 11 and 12: Flat surface mountain chains west of the Quebrada de Sepulturas at ~4,600 m.a.s.l. (left) and east of the Quebrada de Tumbaya at ~2,800 m.a.s.l. (right). This type of flat topography is being interpreted as a structurally and climatically inherited form.

An exhumed paleosurface of Triassic age is known from the Sierras Pampeanas (COSTA ET AL. 1999). The Sierras Pampeanas show a rather different style of basement uplift and should therefore have experienced a different structural evolution (COSTA ET AL. 1999, KLEY ET AL. 1999). Remnants of paleosurfaces have been mapped, described and dated in southern Bolivia between 2,200-3,800 m.a.s.l. and have been interpreted with regard to plateau uplift mechanisms, timing, and rates (GUBBLES ET AL. 1993, KENNAN ET AL. 1997, KENNAN 2000). Due to a similar regional structural framework (KLEY ET AL. 1999), the planation surface which the mountain chains of the Quebrada de Purmamarca have been uplifted from, is likely to correspond to the paleosurface mapped in Bolivia. Against this background, the topography of the flat mountain tops within the study area could be the last relict of a Miocene planation surface, the last evidence of a pre-Andean topography.

The fact that these surfaces have not been dissected and disintegrated by erosional processes in these high-mountainous environments certainly implies several things. The general erosional state of the Cordillera Oriental can be inferred to be relatively young. This points to absent or weak glacial processes, which are known to efficiently erode mountain belts (OLLIER AND PAIN 2000). Instead, the dominant geomorphic processes must have been processes preserving and pronouncing the flat and smooth topographical aspect of the mountain chains.

4.2. DRAINAGE PATTERNS

The overall drainage network of the study area is characterized by rectangular patterns. Particularly, the orientation of the larger lateral valleys follows the clear N-S and NE-SW trend (e.g. Quebrada de Lipán, Sepulturas and Estancia Grande). In contrast, the main valley of the Quebrada de Purmamarca runs approximately W-E to NW-SE. However, the dominating pattern throughout the Cordillera Oriental is an N-S orientation following the structural preconditions. Most valleys have accommodated in fault-lines, zones of structural and geomorphological weakness. The fact that the Quebrada de Purmamarca seems to cut most of these structural features could therefore point to two contrasting implications. In one case, the Quebrada de Purmamarca could be considered an example of an antecedent valley which has inherited and maintained its course from the initial phase of Andean uplift; fluvial incision has always been efficient enough to outpace uplift rates. In the other case, the orientation of the valley might have been supported by a major fault line with W-E orientation.

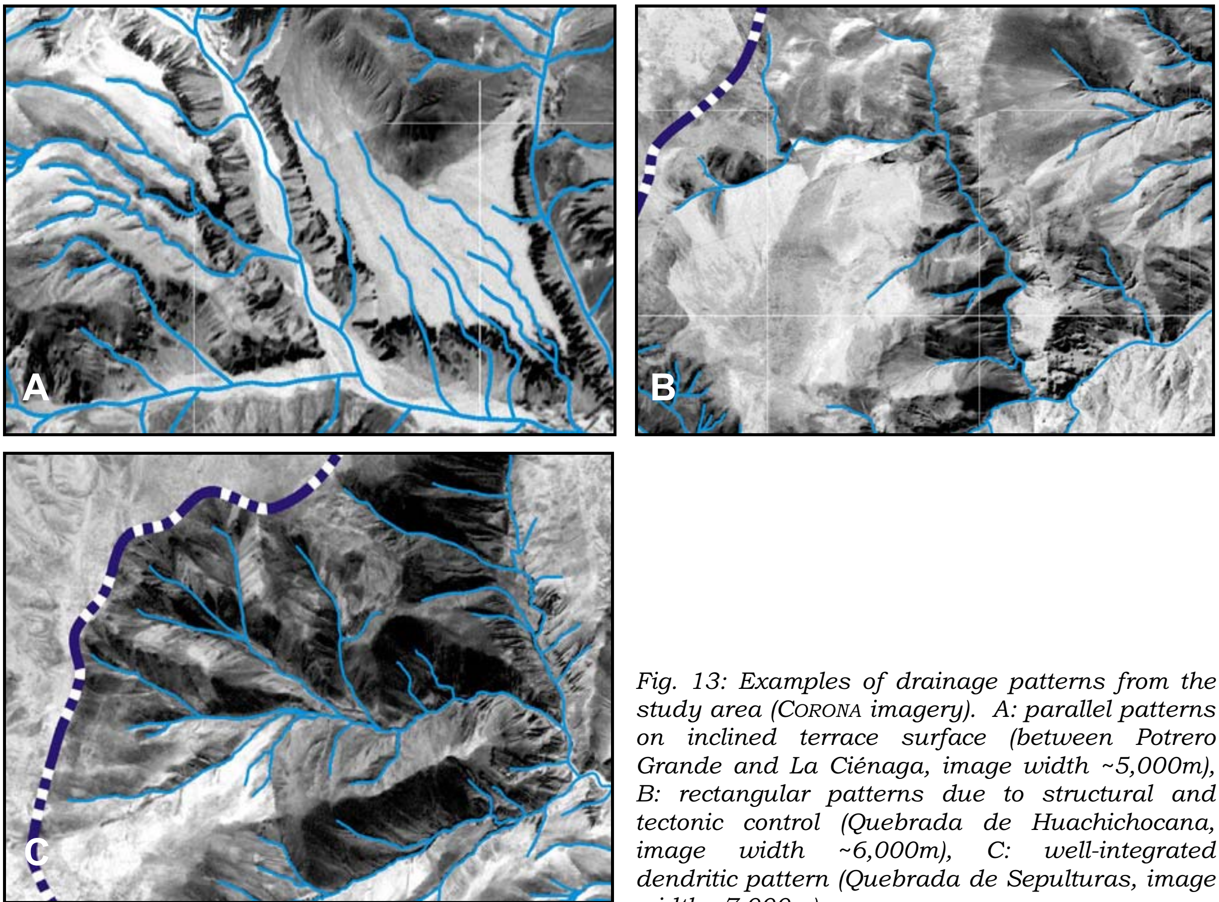


Fig. 13: Examples of drainage patterns from the study area (CORONA imagery). A: parallel patterns on inclined terrace surface (between Potrero Grande and La Ciénaga, image width ~5,000m), B: rectangular patterns due to structural and tectonic control (Quebrada de Huachichocana, image width ~6,000m), C: well-integrated dendritic pattern (Quebrada de Sepulturas, image width ~7,000m).

Fault lines rectangular to the overall structural orientation are common features in the Cordillera Oriental. Although famous examples from Bolivia's Sierras Subandinas (e.g. KENNAN ET AL. 1997, BLOOM 1998) have shown that drainage lines have not changed essentially in some places since late Tertiary times, antecedence can not be tested within

the study area. Considering the enormous Cenozoic uplift rates which have been reconstructed for the Central Andes by several authors (KENNAN ET AL. 1997, GREGORY-WODZICKI 2000, HARRIS AND MIX 2002), it seems likely that a major fault line has essentially controlled the overall drainage evolution of the Quebrada de Purmamarca.
Similar conclusions have been drawn for the entire drainage system of the Quebrada de Humahuaca that has certainly been modified significantly in the geologic past by uplift and subsequent epigenesis (SEGEMART-ITGE 1998).

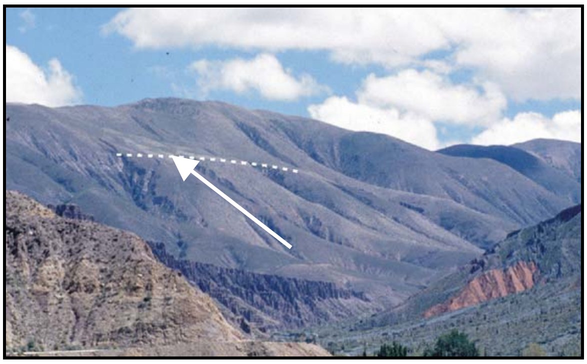
On a larger scale, several deviations from this rectangular drainage pattern have been observed which all reveal local structural and lithological controls. On the relatively young, gently inclined slopes of the larger terraces a system of parallel drainage has developed following the dip. On the lower terraces it has developed a nearly ideal pattern while in the upper terraces it reveals signs of branching at several places (Fig. 13 A).
At other places, quasi-dendritic patterns have evolved. This is particularly true for the eastern slopes of the higher mountain chains (e.g. Quebradas de Sepulturas, Estancia Grande, Chalala, Coqueña; Fig. 13 C). Here, the evolution of the drainage network has most probably experienced better conditions due to a higher amount of precipitation resulting from the eastward aspect of the slopes.
Regarding the shape of the drainage basins, significant differences between the sub-basins of the study area can be observed. While many basins show a rather narrow shape with an approximate N-S orientation corresponding to the regional structural style (e.g. Quebrada de Lipán), the Quebrada de Huachichocana reveals a relatively large and rounded basin shape which could imply an older age or a different tectonic evolution.

4.3. FLUVIAL TERRACES

Fluvial terraces are among the most commonly used features for the reconstruction of landscape history. They are easy to map from most remote sensing data and have been a major focus of field work. Within the study area, they belong to the most striking features, some of them reaching extents of several km^2. By definition fluvial terraces are the relict parts of ancient floodplains. They are cut by fluvial processes but are not necessarily built up from fluvial deposits (e.g. BLOOM 1989). Both types of fluvial terraces, aggradational and erosional terraces, have been distinguished in the study area. Aside from their morphological description, their sedimentological characteristics will be particularly essential for further interpretation. Therefore emphasis will be put on the sedimentological analysis of the terrace deposits.

4.3.1. EROSIONAL TERRACES

At least at three locations within the study area, pronounced bends in slope angle have been observed cutting into bare rock (Fig. 14 and 15). Mostly, these erosional surfaces are of very narrow extent. In addition, they can nowhere be traced laterally over more than several hundreds of meters to one kilometer. Therefore they are very isolated features. They have been observed at different locations and elevations of the study area (2,800 m.a.s.l. in the Quebrada de Coqueña, 3,150 m.a.s.l. in the Quebrada de Purmamarca above Ciénaga, 3,400 m.a.s.l. in the Quebrada de Estancia Grande). Usually, their position is parallel to the course of the present valley.



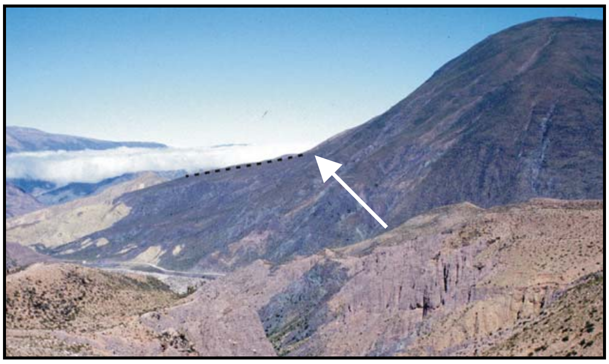


Fig. 14 and 15: Erosional surfaces interpreted as erosional (strath-) terraces in the Quebrada de Estancia Grande, 65,55 W and 23,66 S at ~3,200 m.a.s.l. (left); and in the Quebrada de Purmamarca, 65,55 W and 23,71 S at 3,130 m.a.s.l. (right).

Concluding from their morphological characteristics, these features represent erosional terraces resulting from enhanced lateral erosion during stages of tectonic stability. Only because of their pronounced morphological appearance as erosional surfaces they could be mapped at all, and the fact that they are very isolated features makes it impossible to correlate them to each other. However, they could be viewed as the last remnants of earlier stages of valley and landscape evolution, possibly giving hints to significant regional tectonic movements which resulted in vertical incision and consequently in the isolation of these features. Similar observations of erosional terraces within the Cordillera Oriental without chronological interpretation have been reported from WERNER (1984) and TCHILINGUIRIANI AND PEREYRA (2001).

4.3.2. DEPOSITIONAL TERRACES

Much more evident and easier to identify are the depositional fluvial terraces within the study area. The surface area of the large terrace downstream from Lipán (Terraza Grande) has a surface area of more than 3 km². Within the main valleys of the Quebrada de Purmamarca these terraces are the most dominating morphological features.
The terraces are inclined with angles between 3° and 5,5° corresponding roughly to the inclination of the recent floodplain of 3,5° to 4° (Fig. 18). Locally, the terrace walls stand up to ~160 m (Qbd. de Purmamarca upstream the mouth of Qbd. de Huachichocana) and even ~180 m (Potrero Grande opposite Lipán) above the valley floors. By topographic correlation using remote sensing data as well as GPS data three main levels of fluvial terraces have been divided in the study area (Fig. 18).

- The highest and oldest terrace T-1 is located at Potrero Grande opposite Lipán (Fig. 16 and 17). It stands up approximately 60 m above the next lower terrace and is inclined ~5° to the ESE. The terrace is approximately two kilometers long and 500 meters broad, while not all of this area exhibits a flat surface topography. Most of the terrace surface seems to have been subject to intense fluvial incision, drainage channels have cut >40 m deep into the terrace. Even in their upper parts the channels show a clear V-form. On most satellite images and aerial photography this terrace segment exhibits a slightly darker coloration than the lower terraces.

Fig. 16: Terrace levels T-1 to T-3 (T-1, T-2 Potrero Grande, T-3 terraza grande) from Qbd. del Cobre (viewing WNW).

Fig. 17: Terrace levels T1–T3 (T-1, T-2 Potrero Grande, T-3 Lipán) from above La Ciénaga (viewing NNW).

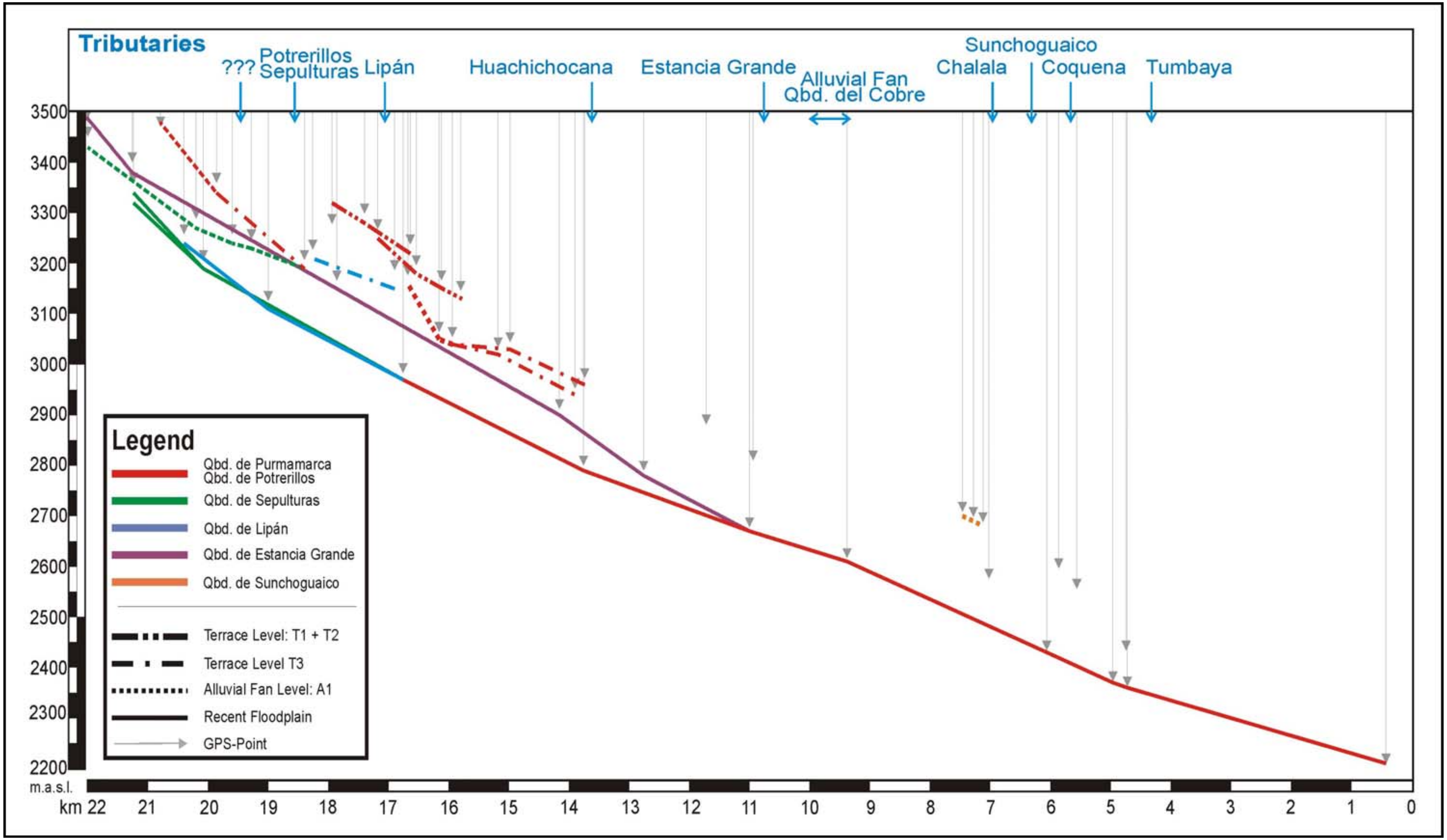


Fig. 18: Longitudinal profile of the Quebrada de Purmamarca and topographic correlation of terrace levels, alluvial fan levels and floodplain using selected GPS-points.

- The terrace T-2 is also located at Potrero Grande just east of T-1 (Fig. 16 and 17). It is inclined ~4,5°-5° to the SSE. Being slightly smaller than the terrace segment of T-1, the surface of the terrace T-2 has an approximate downstream extend of 1,3 kilometers and is 400 meters broad.

The channels cutting through its surface are nowhere deeper than 20 – 30 m, not as wide, and show a V-form as well. In comparison to terrace T-1, the terrace surface seems to be dissected to a much lesser extent.

- The terrace level T-3 has been identified at various locations. Only in the upper reaches of the study area the inclination exceeds 5°, otherwise relatively low inclinations around 3° are characteristic. Several different segments of this terrace level have been topographically correlated to each other. These segments are limited to the Quebradas de Purmamarca, Lipán and Potrerillos. With a surface area of more than six square kilometers the largest segment (Terraza Grande) is located upstream the confluence of the Quebrada de Estancia Grande into the main valley (Fig. 20). Reflecting their lateral position, the terrace segments in the Quebrada de Lipán (Fig. 19) and Quebrada de Potrerillos are much smaller and have an elongated shape. The drainage channels, which have developed on the terrace surface still show the typical steep-walled gully morphology. In addition, they are not as deep as the channels cutting the terrace levels T-1 and T-2. On an average, the surface lies ~90 m below the level T-2, but stands up to ~160 meters above the present valley floor.

Fig. 19: Fluvial terraces (level T-3) at Lipán, height approximately 90 m.

Fig. 20: The "great terrace" (level T-3) north of La Ciénaga, height ~160 m, extend ~2 x 1,5 km.

LABEL	LOCATION	ELEVATION [m.a.s.l.]	DOWNSTREAM EXTENT	INCLINATION	MORPHOLOGICAL CHARACTERISTICS
T-1	Potrero Grande	3,300-3,180 m	1,700 m	5°	V-shaped, deeply dissected
T-2	Potrero Grande	3,240-3,130 m	1,200 m	4,8°	V-shaped medium channels
T-3	Qbd. de Potrerillos	3,190-3,340 m	(1,450 m)	5,5°-6°	Gullies
T-3	Qbd. de Lipán	3,140-3,230 m	2,000 m	2,5°-3,5°	Gullies
T-3	Terraza Grande	2,880-3,040 m	3,400 m	3°	Gullies with steep walls

Table 3: Mapped fluvial terraces and their characteristics.

The terrace segments which have preserved until the present day are all found at locations which has protected them from accelerated erosion. In the Quebrada de Lipán lateral dissection is restricted to a minimum due to the narrow and elongate shape of the quebrada. At Potrero Grande, the terrace segments of generation T-1 and T-2 are protected from lateral erosion due to their elevated position on top of basement rock. The terrace segment of Terraza Grande is situated in a "*ecksporn*"-position between the confluence of three quebradas. However, all of these terrace segments are limited by vertical rims resulting from intense fluvial incision.

Considering the above-described morphological characteristics of the depositional terraces, several preliminary conclusions can be drawn. First of all, the subdivision of three generations of depositional terraces implies a sequence of at least three cut-and-fill events, alternating phases of aggradation and erosion, for the Quebrada de Purmamarca. The enormous size and volume of the terraces point to highly intense aggradational processes, which must have been active during relatively long periods. Particularly the enormous volume and elevation of the terrace generation T-1 can only be explained by contrasting geomorphic processes due to major shifts of environmental or tectonic controls.

4.3.2.1. SEDIMENTOLOGY OF THE TERRACE DEPOSITS

At many places, the enormous vertical exposure of the depositional terraces allows an insight into relict landforms. Therefore, vertical stratigraphical profiles through the steep slopes of the relict depositional landforms like terraces and alluvial fans have been taken at 15 locations (Fig. 21). They exhibit and present evidence for processes and changing environmental conditions responsible for the accumulation of the above-described terraces.

Fig. 21: Typical outcrop (here at Lipán LI-1) of fanglomeratic deposits building up the depo-sitional terraces in the study area (height of the outcrop approximately 40 meters). Note the distinctiveness of the individual layers but uniform overall appear-ance of the fanglome-rates.

SEDIMENTARY LITHOFACIES

The description of the sedimentological profiles is based on the definition of seven different types of lithofacies. As the facies concept relates sediments to their depositional environments, it allows paleoenvironmental conclusions. To some extend the following description and interpretation of the different lithofacies is based on existing classifications for alluvial fan and braided river sediments (NEMEC AND STEEL 1984, BLAIR AND MC PHERSON 1994, MIALL 1996, SOLER AND MAY 2001) because of the overall coarse-clastic character of the observed sediments. Nevertheless, the denomination and definition of the observed lithofacies uses proper criteria depending on local sedimentological characteristics.
The main sedimentological parameters which were used to define the different types of lithofacies are the quantity of matrix (clay to sand-sized material), the assemblage of clasts (texture), the shape of clasts and the average clast size as well as maximum clast size. All of these parameters are easily deducable by field observation from the various outcrop locations along the rims of dissected and eroded terraces. While these parameters form the base for the definition of lithofacies, their nomenclature of D, F and L is largely a result of subsequent sedimentary interpretation.

Lithofacies D1

These lithofacies appear as massive fanglomerates of varying color (Fig. 22). The fanglomerates are matrix-supported, the clasts literally float in a relatively large quantity of matrix and there is no sorting of grain sizes. The average clast size is estimated to be around 10-20 cm while single clasts can reach up to 120 cm in diameter. The clast shape is variable, mostly subangular, and less frequently subrounded. No internal structure is apparent. Usually these lithofacies are built up from strata of no more than 150 cm in thickness, representing single depositional events.
Regarding their lithological characteristics, these lithofacies are interpreted as dominated by deposition of highly cohesive, clast-poor and matrix-supported debris-flows or mud flows accounting for deposition in a medium to distal alluvial fan environment (LOWE 1979, 1982, BLAIR AND MCPHERSON 1994, HARVEY 1997). Alternatively, the low content of clasts might indicate a reduced availability of clasts, possibly resulting from changing environmental conditions.

Lithofacies D2

These lithofacies are the most common and apparently the most representative lithofacies for most deposits in the study area. Like lithofacies D1, lithofacies D2 dominantly appear as massive fanglomerates of varying color. Resulting from the smaller quantity of matrix, the clasts are in close contact to each other, therefore the fanglomerates are clast-supported. There is no overall sorting of grain sizes, but occasionally single strata show a tendency to coarsen upwards.

Fig. 22: Detail of lithofacies D1 (matrix-supported fanglomerate).

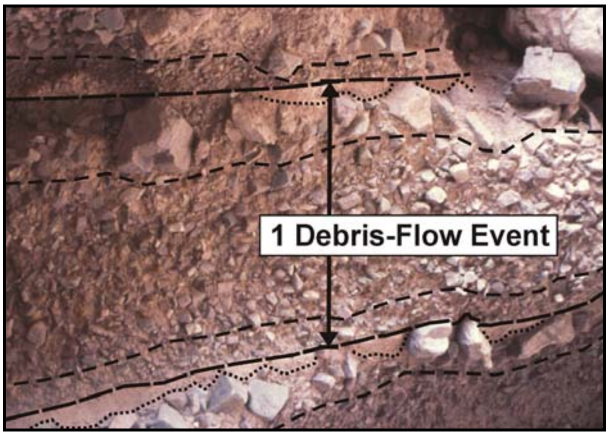


Fig. 23: Detail of lithofacies D2 (clast-supported fanglomerate). Note the internal structure indicating deposition by cohesive debris-flow.

The average clast size does rarely exceed 20 cm and reaches a maximum size of 100 cm. The clast shape varies, but is mostly subangular. These lithofacies are dominated by strata of up to 150 cm in thickness. Very often the strata reveal an internal structure indicative of cohesive debris-flow deposition (Fig. 23). The lowest thin layer is devoid of larger clast, it resembles a traction or inertia layer due to the friction of the ground (LOWE 1982, POSTMA ET AL. 1988). Above that, the thickest part of the debris-flow mass represents the flowing layer of bulk material sometimes called the plug (JOHNSON AND RODINE 1984). It is usually very cohesive and viscous (COSTA 1984, BLAIR AND MCPHERSON 1994). The largest blocks seem to float on top of the other layers. They have moved upwards and maintain their position at the top of the flowing mass mainly due to buoyant, sometimes dispersive or turbulent forces (RODINE AND JOHNSON 1984, COSTA 1984, BLAIR AND MCPHERSON 1998). After cessation of the flow due to dewatering or slope decline the flow mass is subject to subsequent fluvial reworking, winnowing and redeposition of fine material, which forms the lenses and couplets of sand separating the deposits of the individual flow events from each other (BLAIR AND MCPHERSON 1994).

Due to the above-described characteristics these lithofacies can be interpreted as being dominated by deposition of cohesive, clast-rich and clast-supported debris-flows (BLAIR AND MCPHERSON 1994). From their internal structure and the intercalation of sandy layers this type of debris-flow deposition can be interpreted as non-erosive (RODINE AND JOHNSON 1984).

Lithofacies D2X

These lithofacies show many characteristics similar to lithofacies D2 as they have relatively little matrix material and are clearly clast-supported (Fig. 24). There is no apparent sorting, the internal bedding is chaotic. Maximum clast sizes can reach more than 200 cm, while the amount of larger clasts (> 80-100 cm) is striking. Thus the average clast size is estimated to be around 30-50 cm. Usually the thickness of these strata is more than 200 cm, locally more than 500 cm have been observed. The shape of the clasts is dominantly angular, very few clasts are subrounded.

Despite the clear differences in clast size, these lithofacies are interpreted in analogy to lithofacies D2. They resemble deposition by clast-rich and clast-supported debris-flows, even though the quantity and poor sorting of large clast points to dispersion and turbulence as the main supporting mechanism of flow (LOWE 1982, COSTA 1984). The enormous range in clast size seems to give evidence for extraordinary large debris-flow events, possibly related to and triggered by rock slide or rock avalanche events (BLAIR AND MCPHERSON 1994).

Fig. 24: Lithofacies D2X. Note hat at bottom for scale.

Fig. 25: Lithofacies D3. Note the abundance of medium sized clasts.

Lithofacies D3

Lithofacies D3 crops out only at a single sedimentological profile. It can be described as a massive clast-supported fanglomerate without internal structure. Depositional strata have not been observed thicker than 80-100 cm. Average clast size varies between 15-20 cm and maximum clast size does not exceed 30 cm, while clasts exhibit an angular shape. This relatively uniform distribution of clast sizes is characteristic for these lithofacies (Fig. 25).
Therefore lithofacies D3 indicates dominant deposition by non-cohesive, clast-rich debris-flows (BLAIR AND MCPHERSON 1994). As the flow and support mechanisms of this type of debris-flow depends to a great extent on dispersive and turbulent forces rather than on the cohesion between matrix and clasts, it could be regarded as a grain flow type of debris-flow (LOWE 1982).

Lithofacies D4

These lithofacies resemble a fanglomerate composed of medium to fine grain sizes. Clast sizes do not exceed 10 cm, and individual strata are never thicker than 100 cm, in most cases less than 50 cm. They may appear as thin sheets of gravel, sometimes as couplets or lenses. Most clasts are angular to subangular, but rounded clast shapes are not uncommon. Clast imbrication has been observed at some places. Frequently, the deposits show a clear fining-upwards tendency. A coarser basal layer with grain sizes from 5-7 cm grades into a sandy top with weak internal stratification (Fig. 26).

Fig. 26: Lithofacies D4. Note the fining upwards tendency.

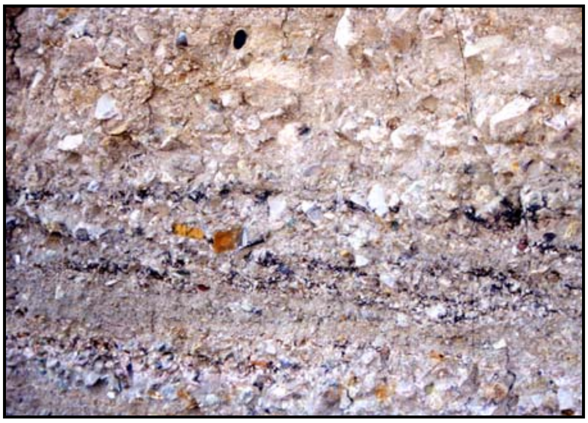


Fig. 27: Thin layer of lithofacies D4 between massive debris-flow deposits. Note black marks, possibly the result of groundwater influence.

Summing up the characteristics of lithofacies D4, these deposits can be interpreted as high-energy waiter-laid sediments in the widest sense (BLAIR AND MCPHERSON 1994, HARVEY 1997). Whether the depositional process occurred as hyperconcentrated streamflow, sheetflooding, sieve or channel deposition is not clear in each individual case. Generally, the fining upwards tendency could indicate sieve deposition while sheets of gravel usually point to sheetflooding or hyperconcentrated streamflow (HARVEY 1997).

Lithofacies F

In contrast to the above-described lithofacies, the deposits of lithofacies F show a clear sorting of grain sizes. In addition, no grains are larger than 4-5 cm, while the bulk of material is sand. Smaller amounts of silty and clayey material are also incorporated and can sometimes form individual strata. Usually the deposited strata are relatively thin, mostly 30-50 cm, even though strata of sand up to 200 cm thick have been observed locally. Very often the thinner strata appear like sheets intercalated between massive layers of lithofacies D1 or D2. The internal structure of lithofacies F ranges from massive to stratified with typical fluvial structure like cross lamination and climbing ripples.

Fig. 28: Lithofacies F, mainly consisting of sand. Note how erosional processes have modelled the internal bedding structure out of the massive sands.

Fig. 29: Detail of lithofacies F. Note the internal structures of cross lamination and local inter-calations of pebble-sized material.

Depending on the appearance, sedimentological characteristics and setting of these deposits, lithofacies F can be interpreted in several ways. It could be the product of fluvial floodplain deposition during a longer time span of decreased debris-flow activity and enhanced availability of sand material. On the other hand, these deposits could be lateral equivalents to the lithofacies L described below, as they have been observed to grade into increasingly finer grain sizes laterally.

Lithofacies L

These lithofacies show the finest granulometry of all lithofacies. Their dominant grain sizes are silt and clay, although sand and very few pebbles do occur locally. Particularly, the good sorting of material and striking internal structures are characteristic for these lithofacies. Bedding is usually horizontal, sometimes concave at the base. Where grain sizes are slightly coarser, structures tend to be more massive and stratification becomes less obvious. Nevertheless, individual beds are never thicker than a few decimeters. The thin beds pile up and form several meters of deposits dominated by lithofacies L (Fig. 30-32). Where clay and silt are the dominant grain sizes, a very fine, flat lamination has been observed. Beds are very thin, sometimes less than a centimeter to a few centimetres, and the lamination shows up particularly well due to color differences within each bed.

Fig. 30-32: Outcrops of lithofacies L in the Quebrada del Cobre CO-2 (left, note yellow-reddish laminated clayed layer at the base, watch for scale), Quebrada de Sunchoguaico SU-1 (middle) and at Potrero Grande PG-1 (right, ^{14}C-date comes from this site).

Color rhythmically changes between a reddish brown and a greenish to greyish yellow. Therefore, these parts of lithofacies L could be described as warves or rhythmites representing cyclic changes, possibly reflecting annual changes in precipitation and sediment input (e.g. TALBOT AND ALLEN 1996). Similar sediments have been reported by TRAUTH AND STRECKER (1999) from several locations in NW-Argentina. Reddish coloration is interpreted to be the result of enhanced sediment input during or shortly after the intense rainy season.

At some places, the flat lamination of the rhythmites is disturbed and bedding planes do not remain in their horizontal position (Fig. 33). This type of convolute bedding might indicate strong bioturbation by plant roots. It seems unlikely that these features have originated from cryogenic deformation or are due to the enormous load of overlying sediments as they occur only locally.

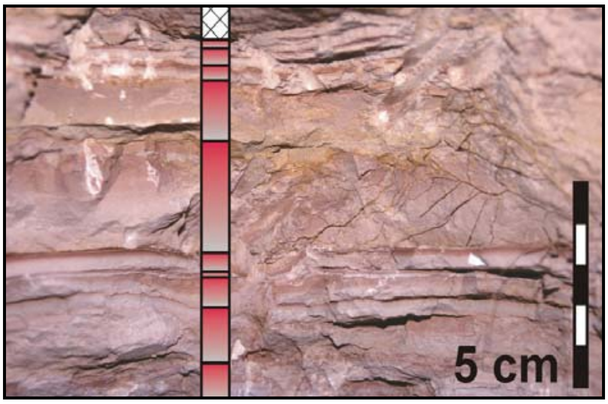


Fig. 33: Fine laminated rhythmites from site SU-1; tendency of cyclic change of coloration within each layer emphasized by schematic bars.

Fig. 34: Layer of convoluted and distorted fine lamination above very silty, pinkish layer. Note undulating base.

Several types of fossils and other evidence for past life has been discovered within the deposits of lithofacies L. Several mm-sized gastropods have been found at various locations, while most of them have been very badly preserved and show up only as imprints. In addition, plant remains and imprints of plant remains have been observed. In one case, the quality of preservation and the amount of remaining organic material were sufficient to extract a radiocarbon date of 49 550 ± 1700 years BP (Appendix). However, the fossils remains have not been determined generically so far.

Summing up the characteristics of lithofacies L, the deposits indicate a low-energy sedimentary environment (e.g. MIALL 1996). The fine material has settled from suspension, while temporal changes in sediment supply were recorded. This is typical for lacustrine environments. The relatively limited occurrence of lithofacies L leads to conclude that the water bodies, which deposited these sediments, were only ephemeral and shallow lakes, not persisting during longer time spans. The question whether these lakes form after blockage of the valley floor induced by landslides or other sediment input (COSTA AND SCHUSTER 1988) or originate from sedimentation in a lateral alluvial fan environment, is an important paleoenvironmental question and will be discussed in the following chapters.

Preliminary Interpretation

Even though the thick accumulations of coarse clastic sediments in the study area have been subject to several publications (FOCHLER-HAUKE 1952, VIERS 1967, WERNER 1984), today no detailed sedimentological description of these sedimentary bodies exist. KÜHN (1924) described them as conglomeratic but mentioned their angular to subrounded character. AMENGUAL AND ZANETTINI (1974) and RAMOS ET AL. (1967) describe them as fanglomerates. Regardless of the denomination as fanglomerates or conglomerates, all authors aim to underline the massive and chaotic appearance of the thick deposits. A first approach to subdividing and interpreting these sediments has been done by SEGEMAR-ITGE (1998) who note them to be conglomerates of fluvial origin. SOLER AND MAY (2001) have analysed four sedimentological profiles at outcrop locations within the study area.
However, on the base of the definition and description of observed lithofacies (table 4), two preliminary conclusions can be drawn. In agreement with earlier geological publications (TURNER AND MON 1979, RAMOS ET AL. 1967, AMENGUAL AND ZANETTINI 1974), the overall classification of the terrace sediments is confirmed to be fanglomeratic. Therefore most lithofacies (D1 to D4) reflect sedimentary characteristics resulting from intense debris-flow activity. Nevertheless, sedimentological and lithological characteristics reveal a relatively large variety and include lithofacies of clear fluvial (F) and lacustrine (L) origin. Even though strictly speaking, the terraces and their deposits are the result of past floodplain aggradation, the proportion of debris-flow deposits is remarkable. The assemblage and co-existence of these lithofacies is commonly interpreted as typical for alluvial fan environments (BLAIR AND MCPHERSON 1994, HARVEY 1997).

	MATRIX	Ø EVENT	CLAST MAX	CLAST Ø	SHAPE	INTERPRETATION	OTHER
D1	+	150	80-120	10-20	Subang. - Subrd.	Debris-flow (~Mud Flow)	-
D2	-	150	100	10-20	Subangular	Debris-flow	Coarsening upw.
D2X	-	> 200	150-200	30-50	Ang. – Subang.	Debris-flow	
D3	-	100	30	15-20	Ang. – Subang.	Debris-flow (~Grain Flow)	
D4	-	< 100	5-10	5-7	Subang. – Round.	Water-Laid	(Fining upw.)
F	/	< 100	Cobbles	1-5	Subang. – Round.	Fluvial	Fining upw. Coarsening upw.
L	/	< 100	Clay - Sand	-	-	Lacustrine	(Coarsening upw.)

Table 4: Summary of main sedimentological characteristics of lithofacies in the study area.

SEDIMENTOLOGICAL PROFILES AND CHARACTERISTICS

Throughout the study area 17 sedimentological profiles from 15 different locations will be described (for overview of profile locations see figure 205 in the appendix). All of the profiles correspond to outcrops, usually along the steep slopes of terraces and alluvial fans. To some extent, the resulting sedimentological information allows interpretation regarding the geomorphic processes during the time of deposition and accumulation. As follows, the main findings and some additional observations add to the information given by the sedimentological columns (Fig. 36-52) and form the base for stratigraphical correlation as well as sedimentological interpretation.

TU-1 The profile TU-1 was taken south of the village of Purmamarca in the Quebrada de Tumbaya. Generally, it consists of fine-grained lithofacies and typical debris-flow lithofacies D2 built up from relatively thin layers. A section of yellowish clayey to silty material rests unconformably on the steeply inclined, orange sandstone bedrock of possible Tertiary origin (compare 2.2.2.). Even though it shows some signs of weathering and erosion, it seems rather resistant and not entirely unconsolidated. In addition, it is dips 17°-18° to the SW and shows several signs of neotectonic movements like microfaults and minor fault striations. This implies a much older age for this section compared to the overlying sections. The sediments are partly laminated and contain few fossil remains (gastropods). Two units of thin layered lithofacies D2, which differ in color, rest unconformably on these sediments. The thin lower unit is orange and the thicker one above is greyish. This seems to indicate different source materials for both of them (orange = Tertiary sandstone, grey = Precambrian schist). There is no soil developed on the top, which leads to conclude that the top of the profile is presently subject to erosional processes.

TU-2 The profile TU-2 was taken west of the village of Purmamarca aside the Quebrada de Tumbaya. It is dominantly built up of lithofacies D2, mostly of a greyish color. The base of these sediments rests unconformably of Cambrian quarzite. In the lower half of the profile fluvial lithofacies F intercalates with lithofacies D2. Towards the top a unit of greyish to pinkish lithofacies D2 is intercalated before grading into lithofacies L. The top of the profile shows no soil and is presently subject to erosional processes. Color differences indicate a changing source area for the profile. Therefore greyish units seem to have received their material from surrounding Precambrian schists.

PU-1 The profile PU-1 was taken in close vicinity to the village of Purmamarca immediately besides the floodplain of the Rìo Purmamarca. Its base is not exposed and it is characterized by interstratified lithofacies D2 and D4. This points to an interaction of debris-flow dominated processes and fluvial processes, similar to the processes observed on the modern floodplain. The overall appearance of the profile is greyish to pinkish, no major changes are detected, which implies a constant source area throughout the entire profile. In the lower part of the profile several black to dark brown streaks and spots have been observed. They originate from weathered clasts. Weathering has possibly been the result of anaerobic conditions close to or within the zone of groundwater. This implies a formerly high local groundwater level in comparison to today, or it simply is result of recent to present processes not related to the sediments. In any case, the profile seems to constantly have maintained a position within floodplain sedimentary environment. There is no soil on top of the profile which leads to conclude that the top part is presently subject to erosional processes.

CH-1 The profile CH-1 was taken aside the Quebrada de Chalala shortly before it leads into the Quebrada de Purmamarca. Its base is not exposed and it is characterized by an assemblage of the various types of debris-flow lithofacies. In the lower and middle part lithofacies L crop out. The top third exhibits a strong color change from greyish-pinkish to yellowish colors. This indicates a change of source material, possible towards a higher content of material received from Ordovician shales which crop out extensively in the Quebrada de Chalala. The top part does not show soil formation. Instead it is covered with accumulated greyish colluvial material originating from the slopes of the surrounding Precambrian schists.

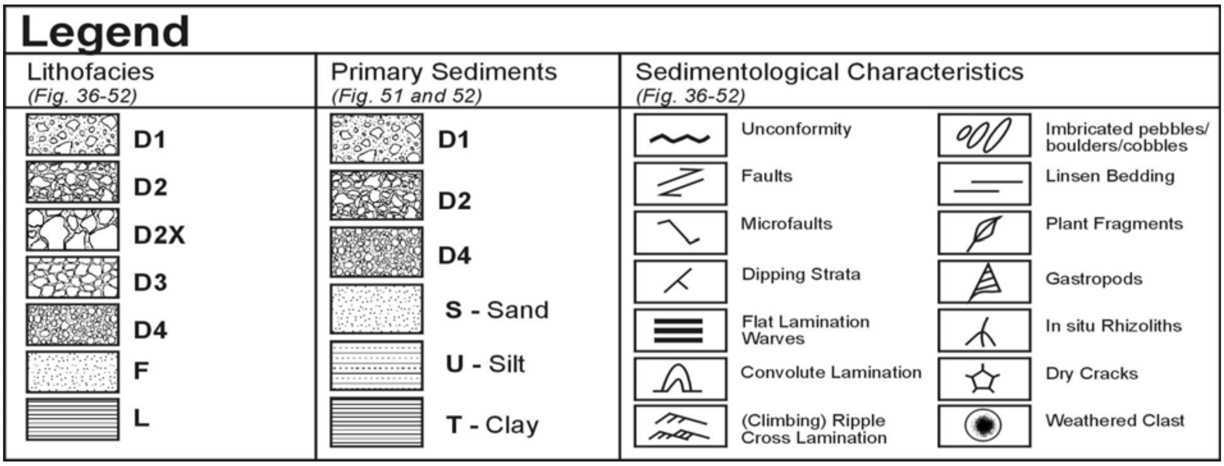


Fig. 35: Legend for the sedimentological profiles (Fig. 36-52)

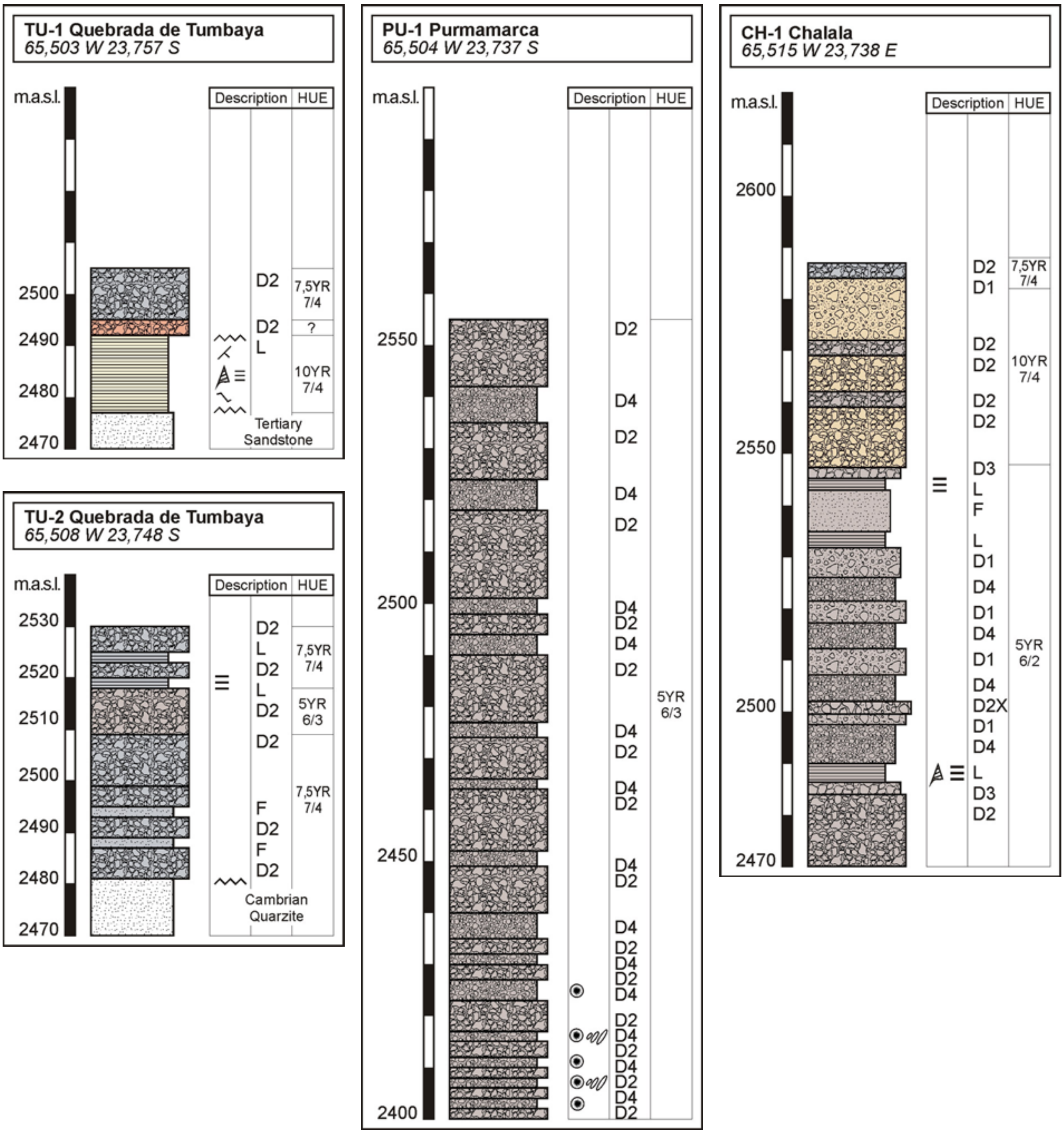


Fig. 36-39: Sedimentological profiles of the study area (TU-1, TU-2, PU-1, CH-1).

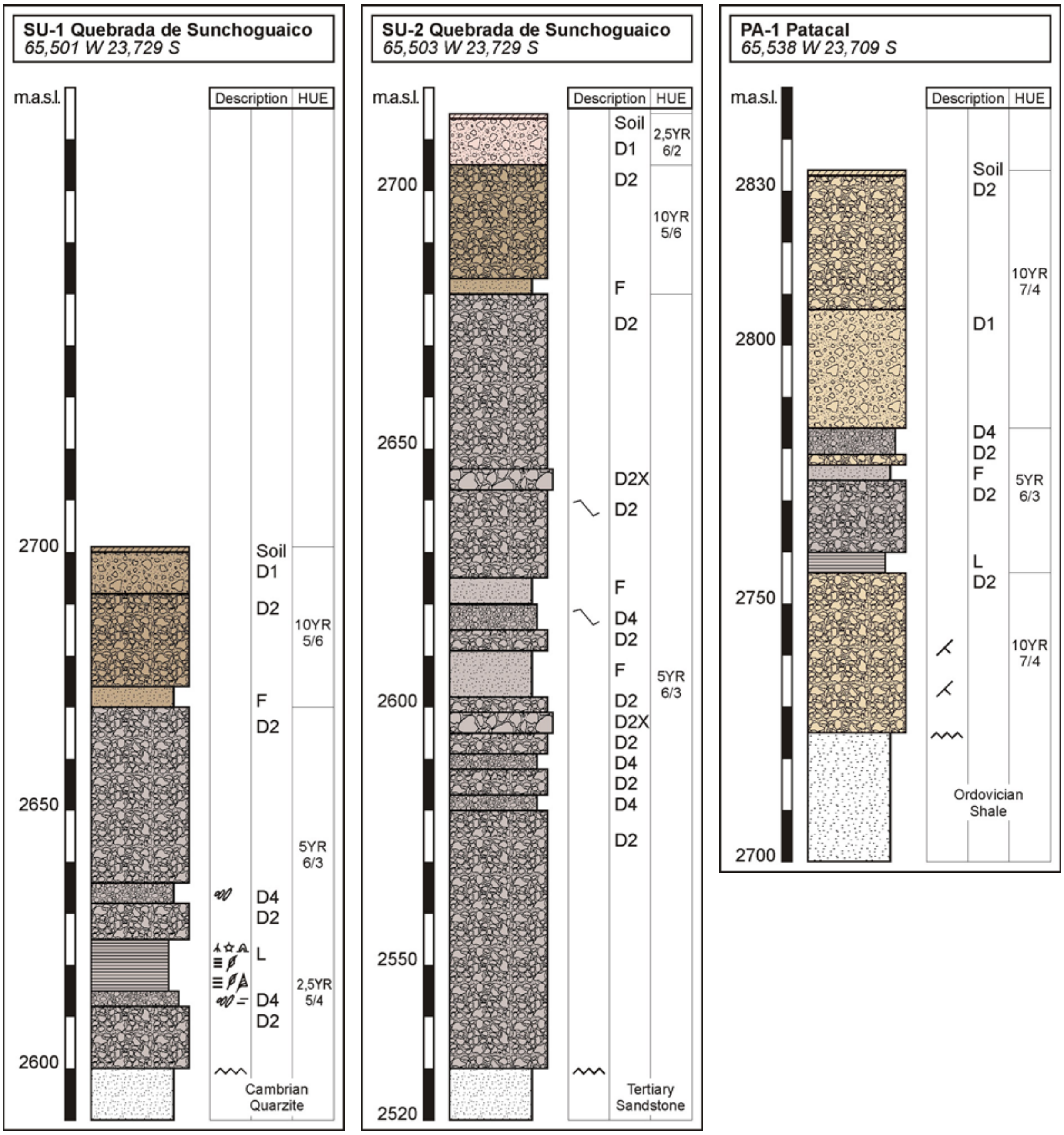


Fig. 40-42: Sedimentological profiles of the study area.(SU-1, SU-2, PA-1).

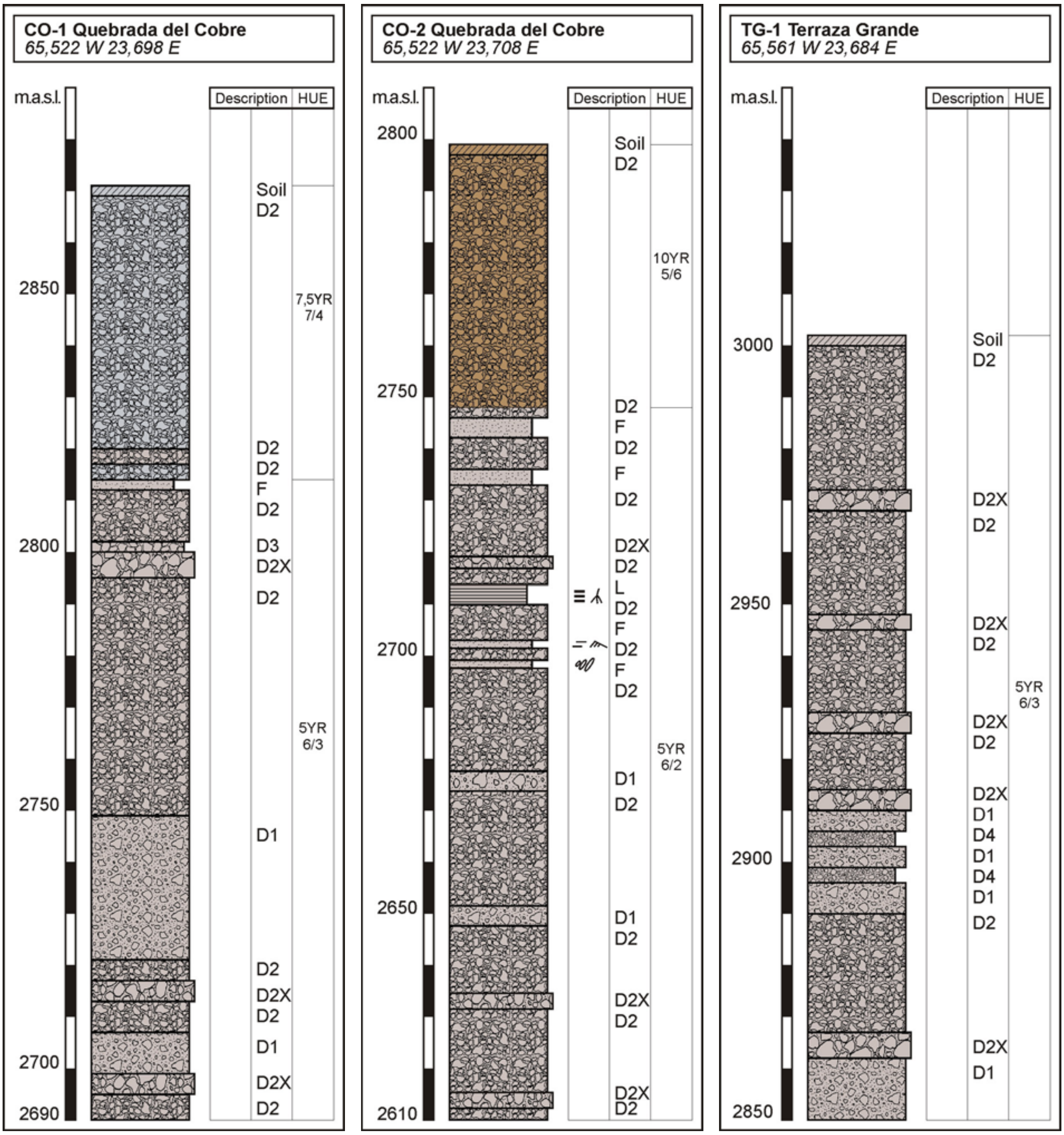


Fig. 43-45: Sedimentological profiles of the study area.(CO-1, CO-2, TG-1).

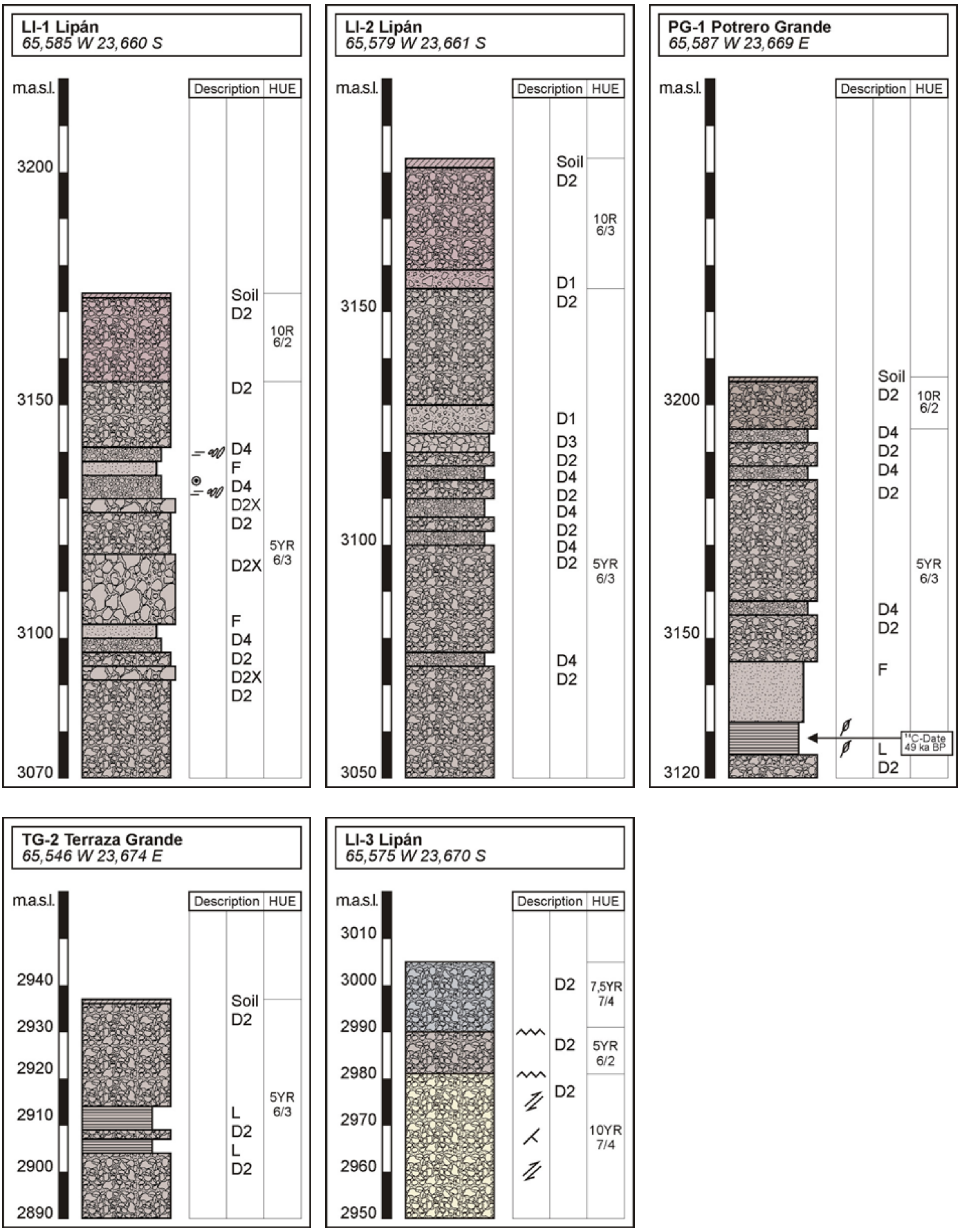
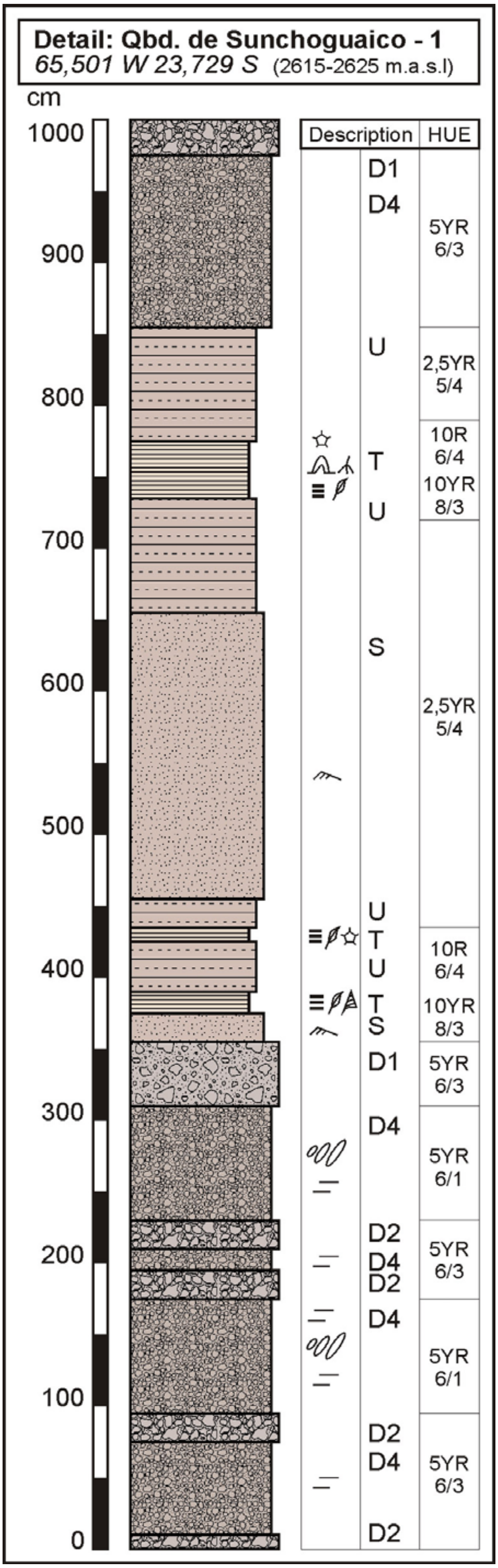


Fig. 46-50: Sedimentological profiles of the study area.(LI-1, LI-2, PG-1, TG-2, LI-3).



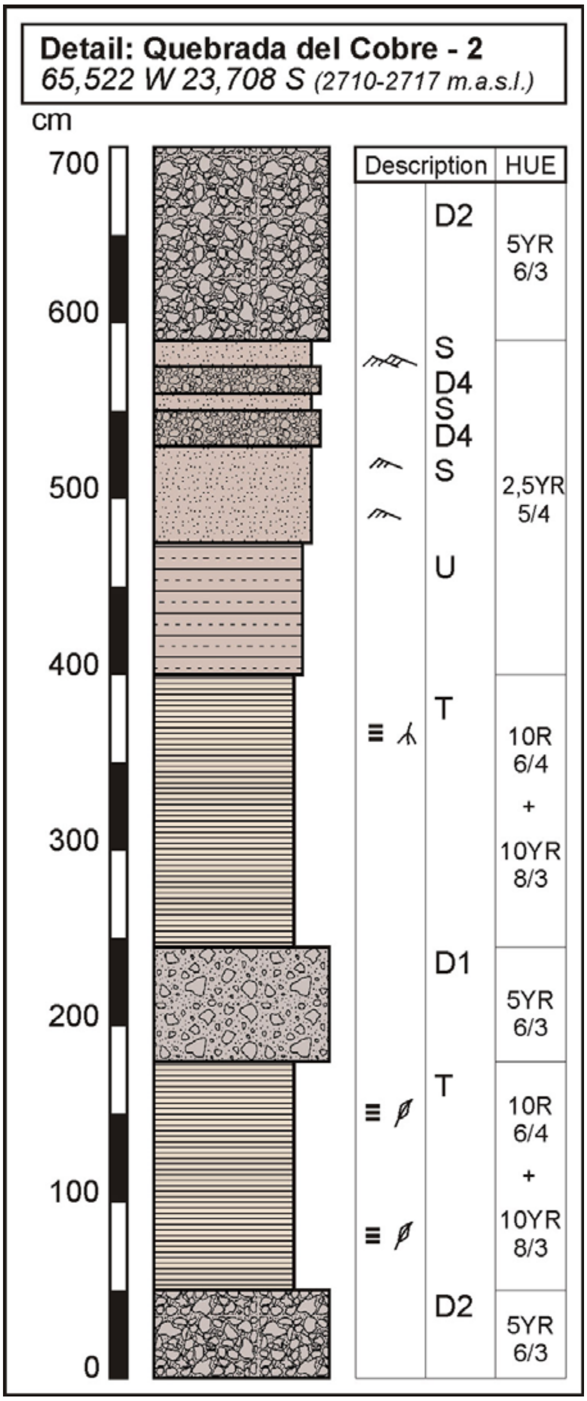


Fig. 51 and 52: Detailed sedimentological profiles in lithofacies L from of the study area.(SU-1, CO-2).

SU-1 The profile SU-1 was taken in the Quebrada de Sunchoguaico. It rests unconformably on Cambrian quartzite and is built up mainly from lithofacies D2 of pinkish to greyish color. In its lower part, a thick unit of lithofacies L contain a variety of fossil remains and give insight into these lithofacies (Fig. 51). The top part shows a brownish color and is separated from the lower part by lithofacies F. The color change indicates a shift in source material towards a higher content of local material like the Cambrian quarzite. A very weak and initial soil has developed on top of the profile.

SU-2 The profile SU-2 was taken in the Quebrada de Sunchoguaico in close vicinity to SU-1. It rests unconformably on Tertiary sandstone and is dominated by lithofacies D2 of greyish to pinkish color. In its middle part two relatively thick outcrops of lithofacies F and lithofacies D2X are noteworthy. In addition, microfaults have been observed at two localities. Towards the top a layer of lithofacies F separates the lower part from two parts of contrasting brownish and pinkish color. More than once a change of source material seems to have taken place and implies a position of the profiles in between two source areas. A weak and initial soil has developed on top of the profile.

PA-1 The profile PA-1 was taken west of the locality of Patacal. In contrast to most other profiles it is dominated by lithofacies D1 and D2 of a yellowish color resting unconformably on Ordovician shales. Only the middle part shows the pinkish-greyish color of most other profiles. In this part there are thin outcrops of lithofacies L and F. The yellowish color might reflect a local source area for most of the deposits of the profile. In the lower part the strata locally seem to dip 15°-24° towards the SW. Whether this is the result of local landsliding or is characteristic for an older, underlying sedimentary unit can not be decided. On top of the profile a weak and initial soil has developed.

CO-1 The profile CO-1 was taken in the Quebrada del Cobre east of Patacal. It is built up almost entirely from the various debris-flow lithofacies and its base is not exposed. The top third of the profile exhibits a color change from pinkish-greyish to greyish source material. This change seems to have taken place gradually which is documented by the intercalation of layers of both colors. On top of the profile an initial to carbonatic soil has developed.

CO-2 The profile CO-2 was taken in the Quebrada del Cobre southeast of Patacal. Its base is not exposed and the profile is characterized mostly by lithofacies D2 of greyish-pinkish color. Towards the top third a color change takes place, the brownish color indicates source material with a higher content of Cambrian quartzite. In the middle part there are outcrops of lithofacies F at several locations and one outcrop of lithofacies L containing very little root material. A weak soil has developed on top of the profile.

TG-1 The profile TG-1 was taken at the western rim of the Terraza Grande opposite the bridge over the Quebrada de Huachichoacana. Its base is not exposed and it is dominated by greyish to pinkish lithofacies D2, but layers of lithofacies D2X crop out at five locations. On top a reddish clay-rich soil has developed.

TG-2 The profile TG-2 was taken at the eastern rim of the Terraza Grande aside the Quebrada de Estancia Grande. Its base is not exposed and it is relatively thin. Like profile TG-1 it is mainly characterized by greyish to pinkish lithofacies D2, in its middle part a massive layer of lithofacies L crop out. A weak soil has developed on top of the profile.

LI-1 The profile LI-1 was taken northwest of the locality of Lipán at the southern rim of a depositional terrace. Its base is not exposed and the profile is dominated by greyish to pinkish debris-flow lithofacies. Particularly striking is the thick layer of lithofacies D2X in the middle part of the profile. In addition lithofacies D4 and F occur, sometimes coupled with streaks of black weathered clasts which point to groundwater influence and floodplain activity during deposition. The top part shows a slight variation in color towards reddish possibly indicating a shift in the area of source material.

LI-2 The profile LI-2 was taken northwest of the locality of Lipán east of the profile LI-1. Its base is not exposed and the profile is characterized by pinkish-greyish lithofacies D2. In its middle part a succession of interstratified lithofacies D2 and D4 points to floodplain activity during deposition. Towards the top the color grades into a slightly more reddish grey. A weak soil has developed on top of the profile.

LI-3 The profile LI-3 was taken immediately besides the floodplain opposite the locality of Lipán. Its base is not exposed. It is entirely dominated by lithofacies D2 while color changes are obvious particularly in the topmost part of the profile. The lower yellowish part is built up from lithofacies D2. Its yellowish color indicates a very different catchment area for source material. Indeed the clast composition of this unit varies significantly to all other profiles as it contains relatively high quantities of andesitic clasts. As there is presently no outcrop of andesite upstream the profile LI-3 severe drainage system changes and a shift in areas of source material must have taken place, or the deposits of this unit are simply old enough, that a formerly overlying andesite must have been completely eroded since the time deposition. Both hypotheses emphasize a relatively old age for this unit in contrast to the overlying units. This assumption is supported by severe faults within the unit. Strata are inclined 24°-28° towards the northwest. On top of this unit a unit of greyish to pinkish horizontal lithofacies D2 rests unconformably on the lowermost unit. The topmost part is formed by a unit of greyish lithofacies D2. As the base of the third unit shows strong topographical ondulations it seems to be separated by the middle unit through an erosional unconformity. The top of the profile does not show any signs of soil formation and is subject to erosional processes.

PG-1 The profile PG-1 was taken besides National Road No.52 south of the Quebrada de Sepulturas. Its base is not exposed and the profile is dominated by pinkish to greyish lithofacies D2. In its lower part a thick succession of lithofacies L and F crops out. It contains plenty of organic material from plant remains. From this layer a radiocarbon date of 49 550 ± 1700 years BP has been extracted (appendix). In most of the profile interstratified layers of lithofacies D4 indicate a position in floodplain vicinity. Towards the top the color grades into a slight reddish grey. On top a weak and initial soil has developed.

Det. SU-1 This profile summarizes a detailed analysis of a section of lithofacies L in profile SU-1 (2,615-2,625 m.a.s.l.). While the lower part is dominated by high-energy water-laid deposits (imbrication) with few interstratified debris-flows, the middle part of fine grained sediments shows two sections of successions from sand to silt and clay. Particularly the second sequence is relatively thick (~400 cm). Clay-rich layers are fine-laminated and contain organic material from fossil remains as well as fossil imprints (molds?). In addition, dessication cracks indicate a shallow and ephemeral character for the lacustrine water body in which these deposits settled. Sandy sections show cross lamination and point to flowing water processes of medium energy. The deposition of lithofacies L finds an abrupt end through the deposition of high-energy lithofacies D4 that re-establishes a debris-flow dominated floodplain depositional environment.

Det. CO-2 This profile summarizes a detailed analysis of a section of lithofacies L in profile CO-2 (2,710-2,717 m.a.s.l.). The lower part consists of very fine-grained clay-rich rhythmites. They follow without transition on top of lithofacies D2 and are interrupted by a thin debris-flow layer. These fine-laminated sediments contain very few plant remains. In a coarsening upwards sequence a succession of silty, sandy and pebbly sections of lithofacies F and D4 sets the stage for the re-establishment of debris-flow dominated depositional environment. While sandy sections show clear signs of cross lamination and climbing ripples indicating an increasing fluvial energy, the pebbly sections of lithofacies D4 are only small lenses of sediment laterally thinning out in short distances. This could imply very local redistributions of coarser sediments in a relatively shallow water body.

GRANULOMETRIC ANALYSIS OF TERRACE DEPOSITS

Besides the descriptive analysis of the sedimentological profiles, some quantitative granulometric analysis of cohesive debris-flow lithofacies in two profiles has been done in order to detect changes in debris-flow matrix composition over time (for data see table 7 in appendix). Hereby, a significant change in matrix composition could indicate altered environmental conditions, e.g. increased chemical weathering.

All of the samples show a composition of grain sizes expected from a debris-flow matrix (Fig. 53). They have a high content of sand (45-92 weight-%), some silt (7-45 weight-%) and very little clay (3-20 weight-%). These results are consistent with most other debris-flow research (e.g. BLAIR AND MCPHERSON 1998) and emphasize the fact that very little clay-sized material is needed to enable the support of larger clasts within the cohesive debris-flow matrix.

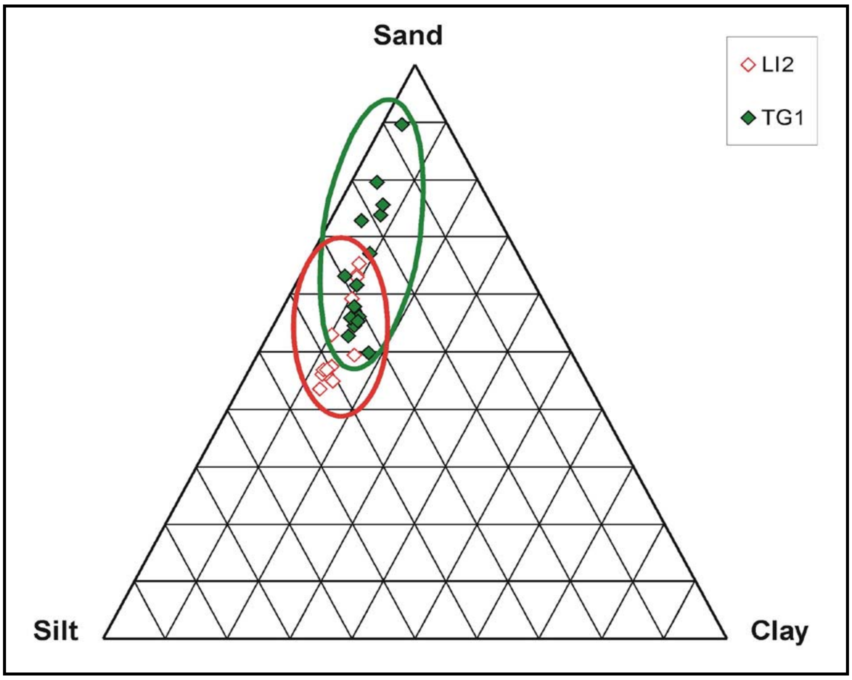


Fig. 53: Grain size distribution of debris-flow matrix samples at LI-2 and TG-1. Note the shift towards coarser grain sizes at TG-1 due to a larger distance to source area, possibly resulting in a process of dewatering and related loss of fine grain sizes.

In addition, a second observation can be made regarding debris-flow transport mechanisms. The grain size samples of the two profiles show different tendencies in grain size composition reflected by the two clusters (Fig. 53). The samples of the profile TG-1 are characterized by a clear shift towards coarser grain sizes, i.e. they contain less silt and clay sized material. This might suggest a relation between the grain size composition of the matrix and the flow distance of each debris-flow, as TG-1 is located much further downstream compared to the profile LI-2. A possible explanation for this differentiation could be the dewatering processes, which the debris-flow is subject to during flowage. The dewatering process constantly removes finer grain sizes from the flow and might eventually cause the debris-flow to cease (COSTA 1984). Because both sampled profiles are associated with the geomorphological generations of terraces T-3, this might lead to the presumption that most debris-flows building up these terraces have one common single source area and have not been brought in by lateral alluvial fan activity.

In both profiles grain size composition shows strong variations reflecting the complex factors controlling the flow mechanics of each individual cohesive debris-flow (Fig. 54 and 55). The variations are within a typical frame for cohesive debris-flow and no overall trend seems visible. The samples of profile LI-2 show a weak tendency towards coarser grain sizes in the top part of the profile. There is no evidence whether this might actually announce a decrease in debris-flow activity and enhanced fluvial activity. In the samples of profile TG-2 this tendency is also visible in the topmost part of the profile, but here it is even harder to interpret it regarding changes of the dominant geomorphic process, as variations occur within a frame characteristic for the entire profile.

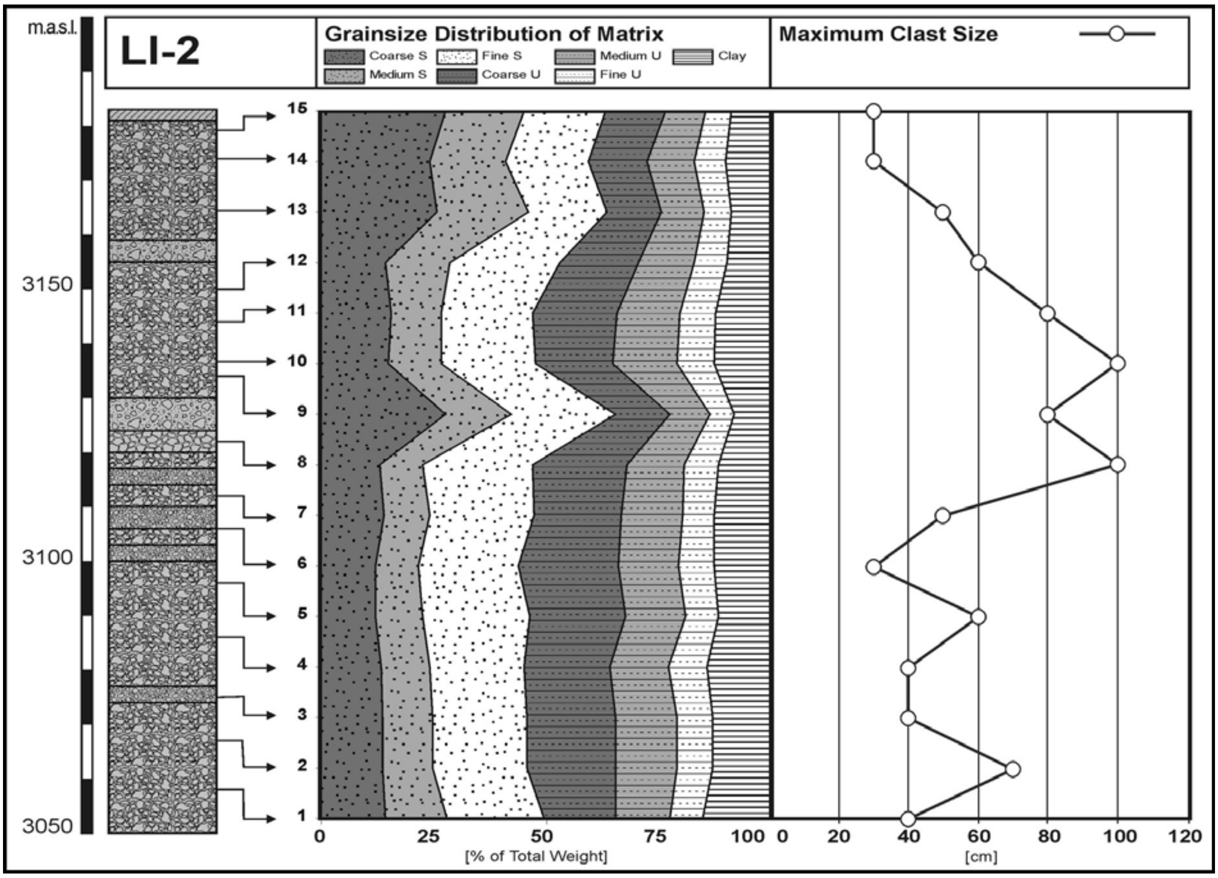


Fig. 54: Grain size composition of cohesive debris-flow matrix at LI-2.

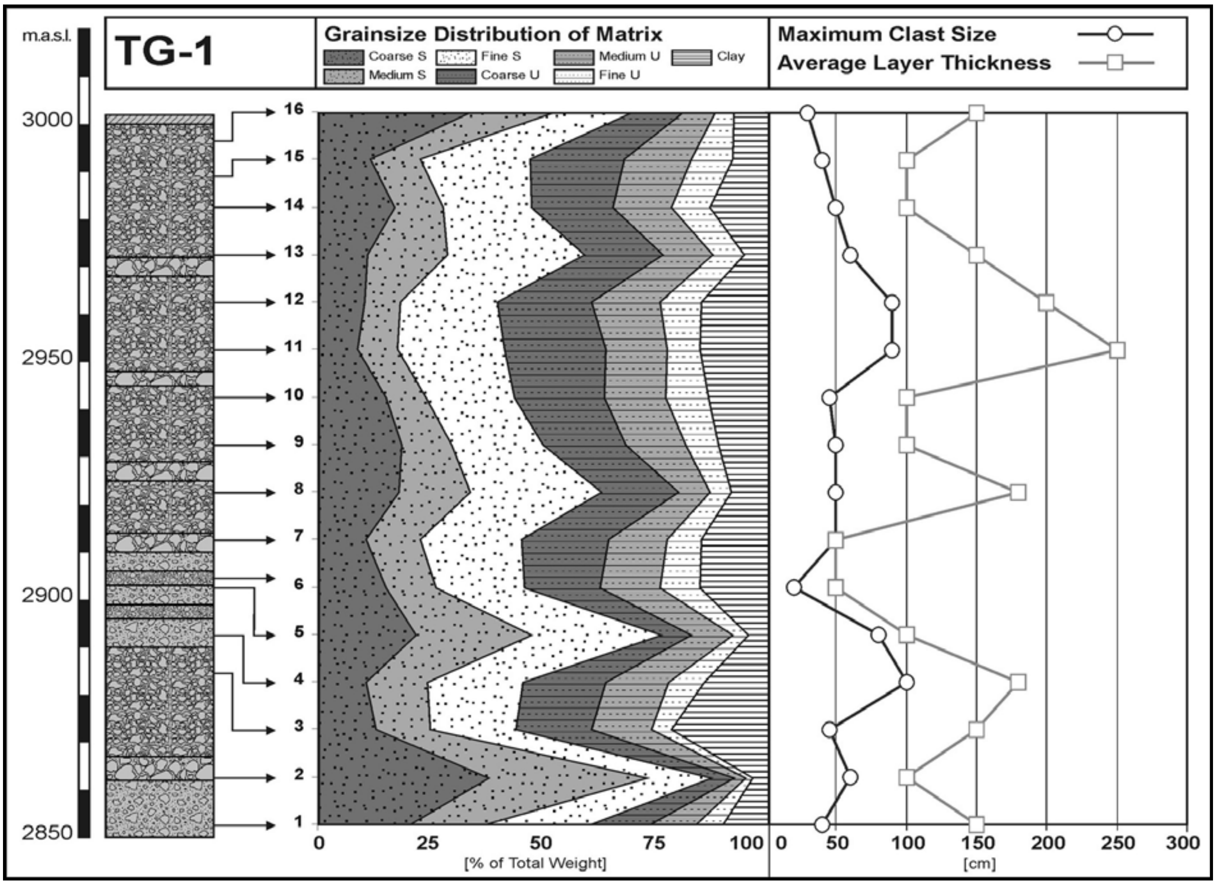


Fig. 55: Grain size composition of cohesive debris-flow matrix at TG-1.

From the analysis of the maximum clast size and the average thickness of the individual debris-flow layers another link can be observed. Both curves do show some variations but have a pronounced minimum where sedimentological and lithological description postulates a section of ceased debris-flow activity and increased fluvial activity. However, the question whether significant changes of matrix grain size composition took place during the period of debris-flow accumulation can only be answered tentatively.

PALEOFLOW DATA

In addition to the above described data paleoflow measurements have been taken at three profiles (LI-1, LI-2, TG-1) in order to obtain information about flow directions and location of main source areas. These measurements used bedding planes of the sandy layers separating the debris-flow deposits from each other. Although inclination and dip direction of these bedding planes will not necessarily correspond to the overall debris-flow direction due to small size topographic differences, a larger number of measurements should give an average trend of paleoflow direction (for data see table 9 in appendix).

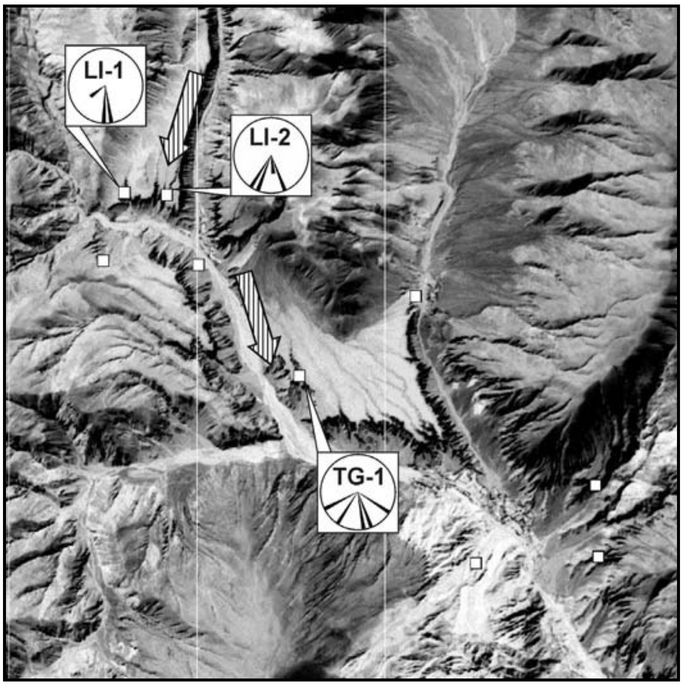


Fig. 56: Reconstructed paleoflow directions for sediments of terrace generation T-3. Note that paleotransport has come from the north, similar to today's valley orientation (CORONA imagery, image width ~9,000 m).

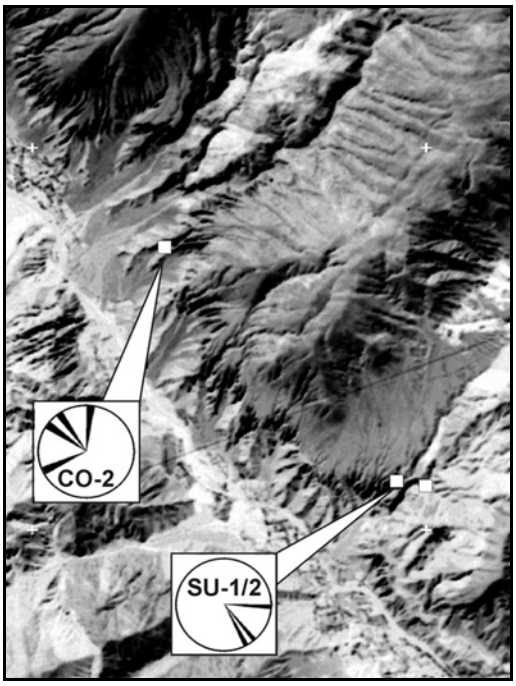


Fig. 57: Reconstructed paleoflow directions for fluvial sediments at two profiles of terrace generation T-3. Note the contrasting directions implying local fluvial transport, possibly towards a paleolake or swamp (CORONA imagery, image width ~5,000 m).

Of course, the obtained results have to be interpreted in a geomorphic context and compared to the present directions of drainage and sediment transport (Fig. 56). LI-2 and LI-1 are parts of the depositional terrace of generation T-3. This terrace section is presently situated at a confluence between two valleys, the Quebrada de Purmamarca coming from the west and the Quebrada de Lipán from the north. However, the measured data imply that paleotransport has exclusively come from the north, leading to conclude that the major source area of T-3 material has been situated in the northern study area between the Quebradas de Lipán and Estancia Grande, where the highest elevations of the entire study area are located (5,036 m.a.s.l.). This observation is consistent with the assumption that the deposition of the terraces of generation T-3 is the result of intense periglacial processes during a cold phase.

Paleoflow directions reconstructed from measurements at profile TG-1 at Terraza Grande show a much larger scatter. Possibly this points to an increasing influence of lateral sediment input from the surrounding slopes with increasing downstream distance. On the other hand, in correspondence to the modern floodplain, the valley floor is likely to have been much wider at Terraza Grande during the time of debris-flow accumulation. As the sandy material separating the debris-flow events from each other mostly results from fluvial processes and secondary winnowing, the bedding planes of these layers are much more likely to reflect small-sized topographic irregularities of the floodplain. Still the average paleoflow direction at TG-1 corresponds approximately to the present situation and is consistent with the measurements at LI-1 and LI-2.

Paleoflow measurements in lithofacies F (Fig. 57) have been taken in three profiles (CO-2, SU-1 and SU-2). At SU-1 and SU-2 the directions are relatively uniform towards the SE. This does not correspond to a SW direction that would have been expected due to the SW-NE orientation of the tributary valley. Therefore lithofacies F could be interpreted as shore facies of small lakes or swamps. This assumption is confirmed as these lithofacies have been observed to laterally grade into lithofacies L. The sediment was delivered from the NW, possibly indicating the pre-existence of an alluvial fan similar to the present topographic situation and from the W where the floodplain must have been located.

In contrast, the paleoflow measurements in lithofacies F at CO-2 show variable directions from SW to N. Considering the present topographic situation, N and NW directions might indicate transport from the central parts of the floodplain into lateral parts, where conditions for fluvial and possibly lacustrine sedimentation were more favourable. SW directions reflect a paleotransport from the adjacent slopes onto the lateral parts of the floodplain.

STRATIGRAPHIC CORRELATION AND INTERPRETATION

Based on the sedimentological and lithological characteristics of the deposits described above, stratigraphical correlation aims to chronologically order the various layers and compare them to each other. Thus extrapolation from local interpretation to regional conclusions becomes possible. Besides, sedimentological information and topographic data are essential for a schematic correlation of the sedimentological profiles in high mountain environments. As this data is attributed to the profiles in figures 36-53, it has not been included in figure 58, but has nevertheless been considered for the correlation.

From the association of the profiles a relatively *pronounced paleotopography* must have formed the base at least for the last cut-and-fill cycle as some profiles rest unconformably on solid bedrock while others do not have their base exposed. That implies that the last intense phase of downcutting must have eroded to a level below the present valley floor, but the depth and the type of *valley fill is not known* at present.

At two locations strongly deformed sedimentary units are found to unconformably underlie horizontal strata. Although these units show very similar lithological and sedimentological characteristics to most other profiles described, they are inferred to be significantly older. They should predate the last intense *regional deformational phase* related to the Andean orogeny. In two cases the strong tectonic deformation, and in one case striking differences in source material lead to conclude that these deposits originate from a time when the topographic and geologic situation still differed significantly from the present situation. However, geomorphic processes seemed to function in analogy to the present conditions as implied by the similarity of debris-flow lithofacies. Therefore it seems likely that the *paleoenvironmental conditions* were characterized by arid to semi-arid conditions.

While by geomorphological analysis three terrace levels had been detected, the sedimentological and stratigraphical correlation shows evidence for accumulation phases older than T-3 only at one location. This might be due to the fact that deposits of generations T-1 and T-2 have only been preserved in one part of the study area. However, all depositional terraces seem to exhibit horizontal bedding, a fact that rules out major tectonic movements since the onset of their depositional cut-and-fill history. Still the erosional unconformities between the three units of profile LI-3 point to a *relatively complex cut-and-fill history* including the two oldest terrace generations of T-1 and T-2.

Overall description and comparison of all profiles reveal two sedimentary sections differing significantly from each other. The lower section consists of varying lithofacies with a predominance of lithofacies D2 and appears greyish-pinkish in all the profiles. The upper section varies in color from profile to profile, even though lithological differences to the lower section are small. In *analogy to the geomorphological results,* the lower section is attributed to the accumulation of the depositional terrace T-3, while in most cases the upper section can be correlated to alluvial fan deposition.

The deposits of T-3 are predominantly built up from debris-flow lithofacies D2, even though detailed description reveal important *lithological variations* within this section, e.g. the floodplain-dominated units as well as lacustrine deposits. The greyish-pinkish overall appearance of the deposits points to similar clastic material and a *constant catchment of source rocks* throughout the period of accumulation, most probably in the northern study area where elevations are highest.

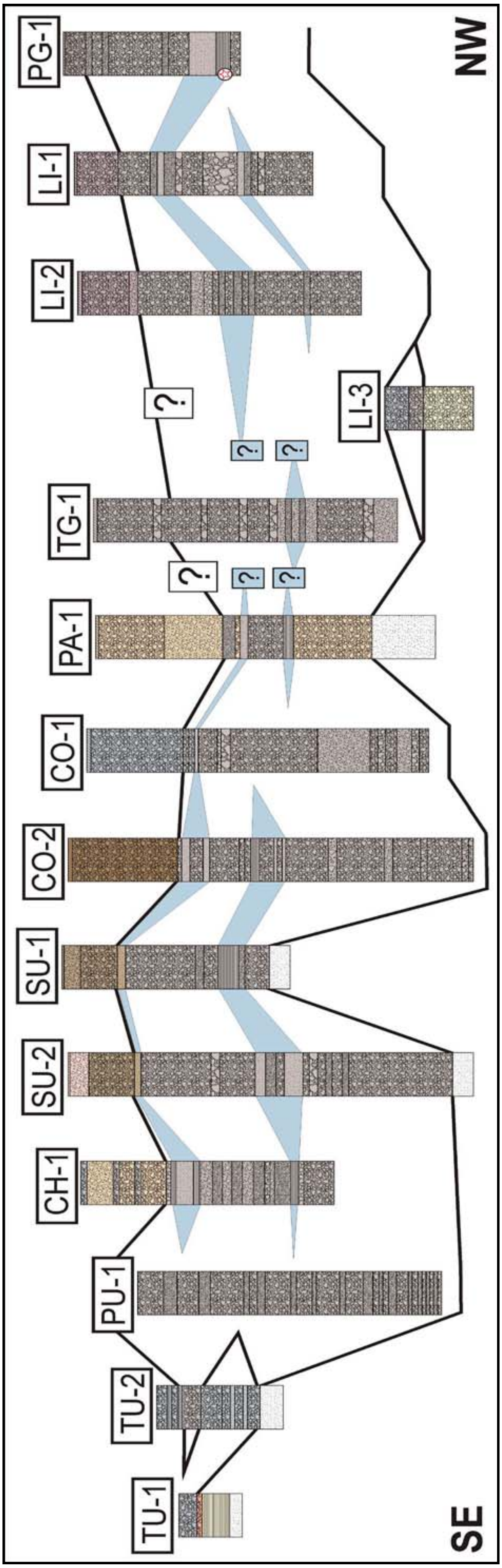


Fig. 58: Schematic stratigraphical correlation of key profiles: cross section Lower Quebrada de Purmamarca (above) and longitudinal section Quebrada de Purmamarca (star marks ^{14}C-date).

The occurrence of lacustrine and/or fluvial sediments seems to be limited to profiles taken in lateral quebradas or in closer vicinity to the steeper valley-limiting bedrock slopes, while they are mostly absent in profiles taken towards the center of the present valley. This might be due to the typical *convex shape* of debris-flow controlled floodplains and alluvial fans (BLAIR AND MCPHERSON 1994, HARVEY 1997), which laterally causes an environment of low depositional energy prone to the damming of lakes and ephemeral swamp-like water bodies. Various examples are known from the present Quebrada de Humahuaca. Lithological and sedimentological description have emphasized their ephemeral and shallow character, suggesting two main *models for their formation* (Fig. 59.).

One of them demands lateral sediment input by debris-flow or landsliding events (COSTA AND SCHUSTER 1988), while the other one explains the lateral water accumulation by lateral damming from an aggradating alluvial floodplain of typically convex shape. For the study area the second model seems more likely because no landslide or other significant deposits have been found in close lateral association with the lacustrine deposits. In addition, the thick and uniform appearance of the T-3 depositional section supports the idea of intense and unidirectional floodplain/alluvial fan aggradation while lateral sediment input was very reduced.

The fact that lithofacies L and most lithofacies F have been observed exclusively within the deposits of section T-3 lead to the conclusion that two contrasting modes of aggradation have been responsible for the deposition of the sections T-3 and overlying section of alluvial fan deposits.

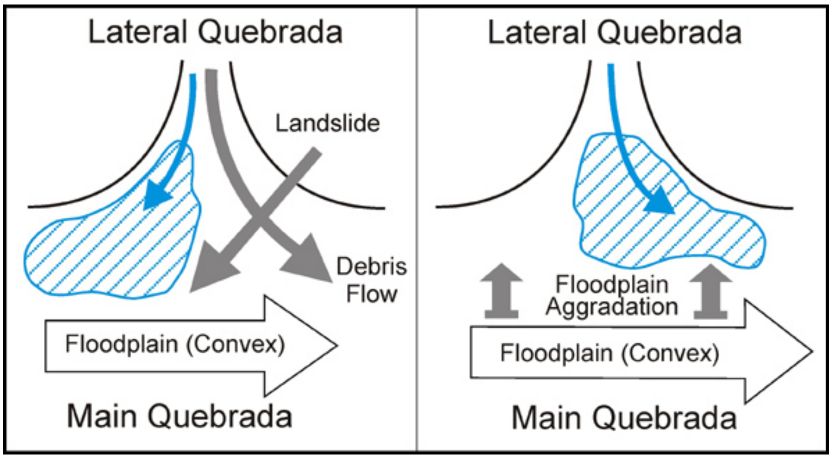


Fig. 59. Two different models for the genesis of ephemeral lakes and swamps in a debris-flow and alluvial fan environ-ment. These types of shallow water bodies or lakes are supposed to be responsible for deposition of lithofacies L and to a lesser extent F.

As evident from figure 59, within the various profiles lithofacies L and F can to some extent be correlated to each other. This should not imply the existence of *one* single interconnected water body because the individual sediments are indicative of shallow water bodies. Considering the rather large distances between the profiles and the high differences in elevation, a much larger lake certainly must have had much greater water depth. It rather suggests the idea of two phases of enhanced conditions for the deposition of these lithofacies. If the model described above is correct, these phases should correspond to a higher influx of water, possibly by increased precipitation. Alternatively, they might indicate

a faster rate of accumulation under otherwise unchanged conditions. However, successions of lithofacies F and D4 within the profiles taken towards the central part of the quebrada or valley floor are consistent with the former explanation and support the idea of phases of *increased humidity* during the period of T-3 deposition.
While both lacustrine phases occur within the section of T-3, the upper one is in many cases directly underlying the uppermost section of alluvial fan deposits and seems to announce a *changing depositional environment* with increasing sedimentary input from lateral small catchments into the main quebrada. The transition from T-3 deposition to alluvial fan deposition takes place gradually. This is particularly evident from localities where greyish-pinkish T-3 material and alluvial fan deposits of contrasting color are intercalated. As the unidirectional debris-flow deposition on the floodplain ceases, lateral alluvial fans start contributing material from many smaller catchments within the Quebrada de Purmamarca. This tendency may also be derived from the grain size composition of the profile LI-2. By a shift towards coarser and sandier matrix composition, the gradually increasing importance of fluvial over debris-flow processes might be reflected.

In the upper study area, on the T-3 terrace surface, a small unit of sediments has been observed with a much higher inclination than the terrace surface itself (Fig. 60). A sedimentological differentiation from the underlying deposits has not been possible, but the directions of its inclination indicate a southwesterly to westerly source of sediment, most likely in the Quebrada de Potrerillos. While terrace deposition had already stopped, this lateral quebrada was still actively producing debris-flows with very similar characteristics to the main terrace level. Neither can it be regarded as part of the underlying terrace deposits, nor does it correlate to the overlying alluvial fan deposits. Therefore it might indicate that the processes responsible for terrace accumulation prevailed longest in the upper study area around the Quebrada de Potrerillos and Sepulturas.

Fig. 60: Onlap of alluvial fan sediments on top of terrace T-3 in the upper study area, implying that fluvial incision eroded the terraces in a retrocedent way.

Even though most alluvial fan deposits in the upper section of the profiles are built up from lithofacies D2 and D1, a clear trend towards thinner bedding and smaller maximum clast size has been observed in most profiles. Corresponding to the smaller catchment area, debris-flow events are notably less catastrophic and intense. Generally, the thickness of the alluvial fan sections is thicker towards the valley sides due to a closer distance to the source area. In addition, the thickness of these deposits correlates to the size and the lithology of the catchment area of these lateral alluvial fans. Deposits from catchments characterized by Ordovician shales show up in the profiles at much lower stratigraphical elevations than deposits from Cambrian or Precambrian catchments. At the same time, larger catchments contribute more material than smaller catchments and build thicker fans. Towards the top of the alluvial fan section, lithofacies D1 get deposited more frequently, possibly an indication for a change of climatic conditions.

PRELIMINARY RESULTS

The above-described results from sedimentological and stratigraphical description and interpretation conclude in the schematic sequence of sedimentary units. Each of these corresponds to an aggradational sequence. Not much can be inferred about the sequence of probable Pliocene to Pleistocene age while the units T-1 to T-3 all seem to correspond to typical climatically induced cut-and-fill phases before conditions change and enable deposition of lateral alluvial fans.

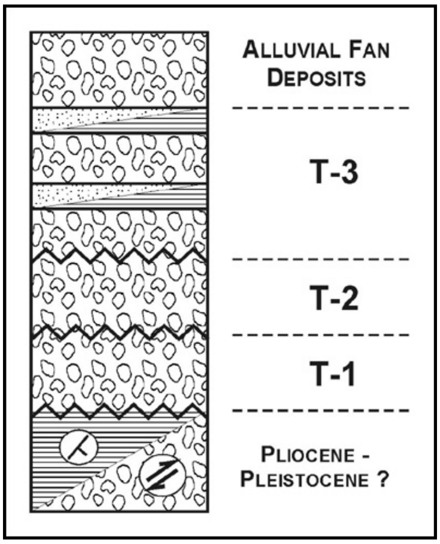


Fig. 61: Schematic sequence of described sedimentary units of the study area. Note different types of unconformities separating the lower units. Each of the units corresponds to a cut-and-fill sequence with the exception of the lateral alluvial fans, which is controlled by local alluvial fan sedimentation.

4.3.2.2. MORPHOLOGY OF TERRACE SURFACES

As mentioned above the terrace surfaces cover relatively large areas within and along the valleys. However, not all of the surfaces show similar morphological characteristics and two contrasting types of surface morphology can be distinguished (Fig. 62 and 63): surface areas of very flat, subhorizontal topography and areas of advanced dissection by drainage channels.

Fig. 62: Terrace segment at Terraza Grande largely characterized by smooth, flat surface area; note very few branches in drainage network (CORONA imagery, width of image ~2,600 m).

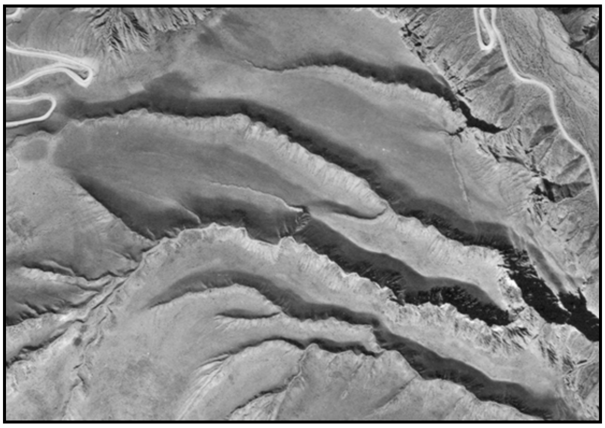


Fig. 63: Deeply dissected terrace segments at Potrero Grande; note branching to almost dendritic patterns in upper parts of the channels (GASATACAMA aerial photography, width of image ~2,450 m).

Most of the terrace surface area is taken up by very smooth plains. Within several hundred of meters up to a few kilometers these plains do not show any topographic irregularity of more than a few decimeters. Their inclination corresponds to the terrace inclination and nowhere exceeds 5° degrees. The very sparse vegetational cover consists of a few isolated shrubs and grasses. Otherwise the ground of these plains is entirely covered by rock clasts, usually ranging from pebble to cobbles sizes. Apparently no fine material is exposed at the surface, a phenomenon commonly described as desert pavement. Depending on the predominant clast lithology, the clast size of the desert pavement varies.

Fig. 64: Desert pavement of schists on alluvial fan in the Quebrada de Sunchoguaico.

Fig. 65: Desert pavement on top of the terrace surface of generation T-3 at Terraza Grande.

The average clast size of pavements on surfaces dominated by schists and phyllites ranges between one and five centimeters, most surfaces with dominant quartzitic lithologies are characterized by an average clast of approximately ten centimeters (Fig. 64 and 65). This seems to reflect the varying lithological resistance to weathering processes.

Despite their exposure at the surface, the clasts do not show any signs of disintegration by chemical weathering. Partly, they are subject to lichen growth, but no type of coating or varnish has been observed. In any case, the occurrence of desert pavements on top of terraces and alluvial fans corroborates the relict and inactive character of these landforms. Furthermore, the desert pavements have important effects on the surface processes and the development of the individual landform. Due to the enormous superficial concentration of clasts, the infiltration capacity of the surface is significantly lowered. Thus, large quantities of runoff are being collected on the extended surfaces of terraces and alluvial fans, particularly as a result of the intense precipitation events characteristic for the semi-arid study area. Moving downslope as overland flow (slope- or sheet wash), the entire water mass reaches the terrace rim, where it constitutes an important input regarding the slope processes shaping the terrace scarps.

In contrast to the above-described flat surface areas, great parts of the terrace surfaces have been subject to fluvial dissection (Fig. 63), even though significant differences in type and intensity of dissection have been observed.

At Potrero Grande deep V-shaped drainage channels have incised into the terrace segments of generation T-1 and T-2 (Fig. 63). On top of the higher terrace segment T-1 drainage channels are incised up to 60 meters into the terrace surface. Channel slopes show inclination angles between 25° and 45° and the entire width of the incised valley reaches up to 200 meters. With a depth of 20 – 40 meters and a width of less than 150 meters, drainage channels on the next lower terrace segment T-2 are significantly smaller.

Fig. 66: Gully on terrace top. Note colluvial material and vegetation on steep gully banks.

Fig. 67: Gully headcut on top of terrace T-3 at Lipán. Note distinct soil horizon of high erosional resisitivity.

Contrary to the V-shaped valley morphology inherent to both terraces T-1 and T-2, the channels dissecting the terrace segments T-3 show very different morphological characteristics. On an average they are not deeper than 10 - 15 meters and 20 – 40 meters wide, while they have remarkably steep walls with slope inclinations of 60° – 90° (Fig. 66). Therefore, these stream channels could be classified as gullies (CAMPBELL 1997), probably representing a first stage in drainage network evolution on top of the terrace T-3. In many cases, their upslope extend is limited by headcuts (Fig. 67), which indicates progressing retrocedent headward growth (THORNES 1994) and supports the idea that gullies are actively evolving on top on the terrace surfaces.

As mentioned above, drainage patterns on top of all terrace generations follow the slope inclination to form longitudinal patterns (4.2.). In addition to the differences in drainage channel morphology, the drainage channel network seems to show significant differences regarding the network evolution. While the drainage network pattern on the surface of terrace T-3 is largely longitudinal, the network on terraces T-1 and T-2 exhibits examples of advanced network evolution. At several places the longitudinal channels branch out in almost dendritic patterns (Fig. 63). In addition, on terrace T-1 examples of drainage capture by present gullying have been observed (Fig. 68).

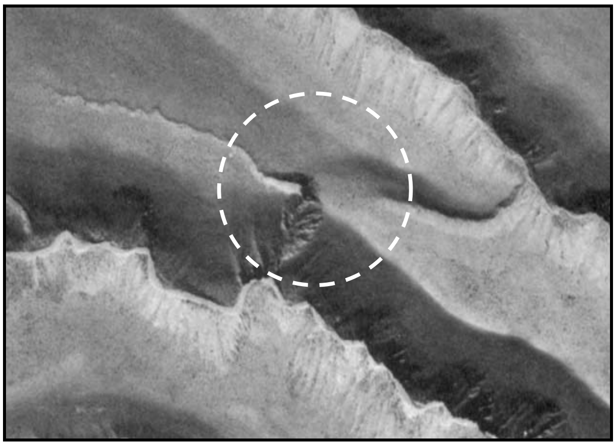


Fig. 68: Drainage capture on terrace T-1 at Potrero Grande reflecting an advanced stage of drainage network evolution (GASATACAMA aerial photography, width of image ~850 m).

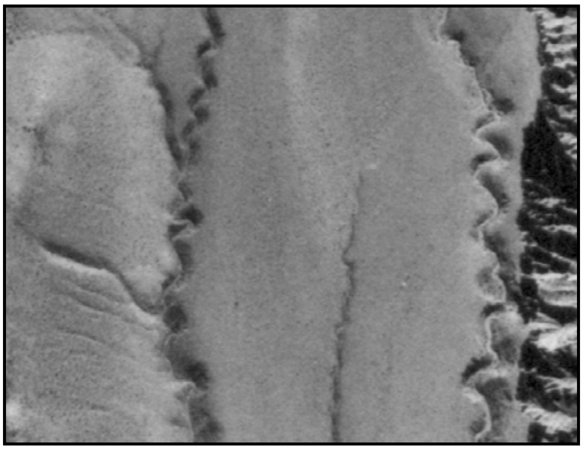


Fig. 69: Meandering drainage channel pattern on terrace T-3 (GASATACAMA aerial photography, width of image ~550 m).

On a larger scale, most drainage channels on top of the terrace surfaces show a relatively straight to slightly irregular channel pattern. On terrace T-3 however, various examples of meandering channel patterns have been observed, e.g. on top of the terrace segments at Lipán and Terraza Grande. While the stream channels have incised up to 15 meters into the terrace surface, sinuosities of up to 1,7 are characteristic for the meandering channel geometry.

Even though no general agreement exists in the literature, the tendency of a river to meander usually indicates a low river gradient, a relatively stable floodplain due to an increased percentage of fine grain sizes (e.g. by intensified chemical weathering or a denser vegetational cover) and lower flow velocities (KNIGHTON 1998). All of these indicators might point to conditions much different, possibly wetter than today, which must have preceded a phase of incision and erosion. The meandering pattern was preserved due to the subsequent channel incision. Therefore the incised meanders might represent two individual phases in the development of the fluvial system of the study area.

Concluding from the morphological characteristics of the terrace surface, two contrasting processes presently dominate the development of the surfaces. While large areas covered with desert pavement are essentially inactive, linear fluvial prosses have partly dissected the terrace surfaces and continue to do so.
By different intensities of drainage network evolution and stream channel morphology, conclusions can be drawn regarding the relative age of the three terrace generations. As expected, the drainage network on top of the highest and likely the oldest terrace T-1 reveals evidence for a rather advanced drainage evolution. In contrast, network evolution as well as channel morphology of terrace T-3 are characteristic for a relatively initial state of dissection. At many places channels still have the morphological characteristics typical of gullies. This can be interpreted as further evidence for significant age differences between the three terrace generations with each of them corresponding to a cut-and-fill cycle.
Nevertheless, drainage evolution on the youngest terrace generation T-3 exhibits evidence for a much more complex evolution of the fluvial environments since the deposition of the terrace T-3. Prior to the intense incision and erosion inherent to each terrace, conditions favoring the development of meandering channel pattern must have prevailed, possibly pointing to wetter conditions than today.

4.3.2.3. MORPHOLOGY OF TERRACE SLOPES

As concluded above, the depositional terraces must have undergone intense erosion. At some places, the terrace walls stand up to 160 meters above the present valley floor, exposing the coarse-clastic terrace sediments. This enormous relief combined with an intense input of water from the terrace surfaces makes these areas vulnerable to a variety of geomorphic processes.

BADLANDS

Probably the most remarkable landscapes characteristic for the lower terrace slopes are the areas of pronounced badland morphology, indicated by a variety of different forms. Within the study area, the badlands virtually concentrate within a belt below and along the rims of the terrace walls. Usually, they occupy an area no wider than 150 – 500 meters, while the vertical extent can reach almost 200 meters. Corresponding to their morphological setting, the badlands have formed within the predominantly fanglomeratic terrace sediments. Their overall morphological appearance is characterized by a highly branched network of small narrow gullies and steep walls (Fig. 70 and 71).

Fig. 70: Badland formation in the Quebrada de Sunchoguaico. Note color variation towards the top indicating a change in catchment lithology.

Fig. 71: Badland formation and "gothic morphology" in Quaternary fanglomerates close to La Cienága.

Many of the walls reach up to the rim of the terrace surface. Very often their inclination angle exceeds 60° - 70° with some of the walls being subvertical. Clay coatings have been observed to cover parts of the walls. In addition, most walls are divided into funnel-shaped vertical gullies or tubes of several tens of meters height (Fig. 72). These features have prepared almost vertical pinnacles out of the wall. FOCHLER-HAUKE (1952) and KÜHN (1924), who have mentioned these features as characteristic for regional badland morphology, have referred to them as *"organ pipes" (German: "Orgelpfeifen", KÜHN 1924)* or *"gothic morphology" (German: "Gothische Morphologie", KÜHN 1924).* Most probably, these features are the result of sheetwash processes on the terrace surfaces. Once the sheetflow has reached the terrace rim, it flows and falls down the vertical walls, each time washing a little amount of the easily soluble fine-grained matrix out of the terrace deposits. The clay coatings are a direct result of redeposition of this transported load, but much of the fine grain sizes is likely being evacuated and removed from the terrace walls, a process essentially enhancing the badland formation. Where large boulders are incorporated in the terrace deposits, they may protect the underlying sediment from wash processes and thus leading to the formation of earth pyramids (Fig. 73). BECKER (1966) studied this type of features and noted them to be typical features of active badland formation under semi-arid conditions.

Downslope, the vertical, funnel-shaped tubes typical for the above-described *"gothic morphology"* transits into deep and narrow gullies. The gully can reach a depth of several tens of meters, while it may be less than a few meters wide. Usually, it is bounded by steep walls. Locally, accumulations of blocks of several meters in size have been observed within the gullies. In most cases, they could be associated with scarps in the walls. These blocks are supposed to result from wall collapse due to lateral undercutting during the seasonal floods. Within the branched network of gullies, a large quantity of water collects and moves downslope confined to the narrow gully channels. Accumulation of collapsed blocks within the gully may cause temporal blockage but generally the collapsed material gets incorporated and removed readily by the intense runoff. This process has been reported as a potential trigger for debris-flows because of the dense mixture of water with sediment of all grain sizes (BLAIR AND MCPHERSON 1994).

Fig. 72: Viewing up an almost ver-tical gully (organ pipe) of approxi-mately 2 m Ø and 50 m height.

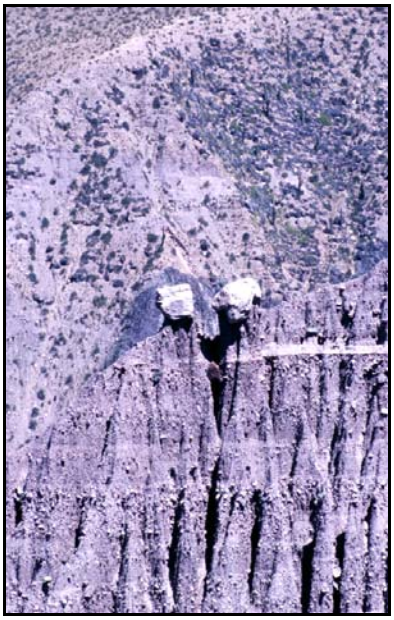


Fig. 73: Larger boulders of 1,5 m Ø forming earth pyramids by protection of underlying sediment.

Fig. 74: Blocks of unconsolidated fanglomeratic material resulting from wall collapse in badland areas. Note potential for debris-flow initiation.

In most cases, the downslope transition from badland areas to the present valley floor is indicated by the deposition of debris-flows forming alluvial fans. These fans do not have a radial length of more than a few hundred meters. Contributing material to the floodplain, these small alluvial fans are frequently subject to lateral fluvial erosion.

LANDSLIDES

As described above, the horizontal bedding of the terrace deposits is particularly well visible within the extended areas of badland formation. Nevertheless, at some places bedding of the deposits has been observed to deviate from its original horizontal position (Fig. 75). Bedding appears to be inclined between 15° and 40°, in all cases dipping towards the slope and away from the quebrada. Showing a sharp transition to the surrounding horizontal deposits, the inclination of bedding is apparently restricted to blocks of several tens until a few hundreds of meters extent. The base of these tilted blocks is nowhere exposed. Usually it is mantled by alluvial deposits of the present floodplain (Fig. 75).
Examples for these tilted blocks of terrace deposits have been identified in the Quebrada de Estancia Grande (~ dip 15° to the NW), the Quebrada de Chalala (~ dip 20° - 40° to the SE) and close to Patacal (~ dip 24° to the SW, Fig. 76). Concluding from their morphological characteristics these blocks of tilted terrace deposits have been interpreted as landslides. The inclination towards the slope and away from the quebrada is typical for the rotational downhill movement along shear surfaces (HANSEN 1984).

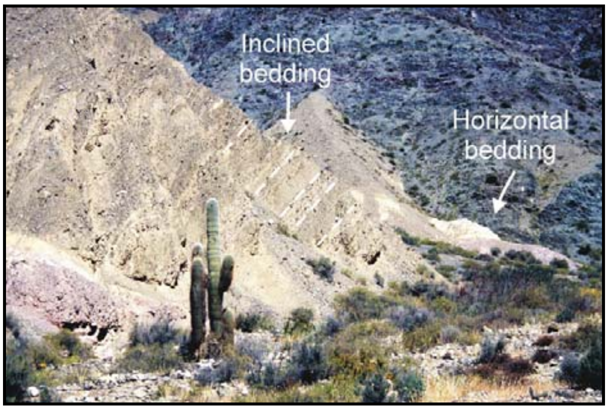


Fig. 75: Steeply Inclined bedding planes of the sliding mass in the Quebrada de Chalala.

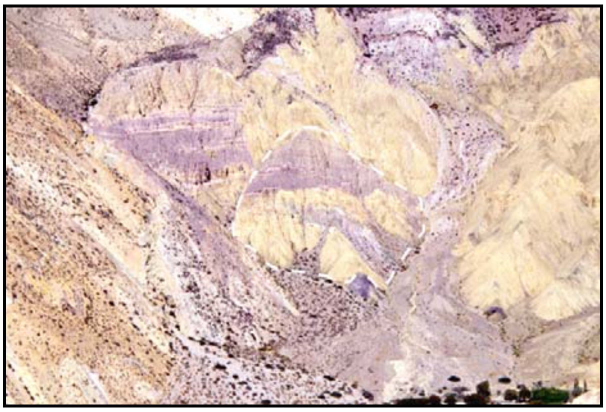


Fig. 76: Landslide in Quaternary fanglomerates (sliding mass marked by dotted line).

While in two cases (Quebradas de Chalala and Patacal) the landslide occurred in association with fine-grained terrce deposits (lithofacies L and F), the landslide observed in the Quebrada de Estancia Grande comprised exclusively coarse-grained lithofacies D1 and D2. Nevertheless, fine-grained lithofacies L and F may have served as sliding planes in most other landslides.

In any case, the landslides within the badland areas seem to be relict features as they are mantled by alluvial sediments. In addition, no landslide scars could be identified, possibly because morphological evidence has been blurred by subsequent fluvial processes inherent to the badland areas. Most likely the landslides have occurred during a time of enhanced conditions for mass-movements. This might point to a phase of increased amounts of humidity, lowering the shear strength of the terrace slopes due to higher moisture content of the sediments. On the other hand, vertical incision and lateral undercutting during a period of more efficient fluvial processes than today could have lowered the slope stability significantly. Finally, neotectonic activity and earthquakes have enormous effects on slope stability as well and might have triggered the observed landslides.

COLLUVIAL SLOPES

While in most parts of the Quebrada de Purmamarca, areas of dominant badland morphology directly pass over into the floodplain, at some places steep hillslope forms have been observed to mantle the lower parts of the badland areas (Fig. 77). These forms have concave slopes of typically 35° - 45° and consist of very coarse and poorly sorted, weakly stratified, matrix-poor debris. Their downslope extent is always less than 100 meters, while they may show a remarkable lateral continuity of more than one kilometer. This type of slope morphology has been observed particularly in the upper study area independent of exposition, e.g. in the Quebrada de Lipán and Purmamarca. Due to their inherent linear character of deposition, fluvial and alluvial processes are not likely to have formed and deposited these forms.

Fig. 77: Relict colluvial slopes in the Quebrada de Lipán; note vegetational cover, erosional scarp from floodplain activity and dissection by gullies.

Fig. 78: Fluvially dissected colluvial slopes in the Quebrada de Purmamarca close to Lipán; note the isolated slope remnant highlighted by circle.

Considering the above-described morphological characteristics, this type of laterally continuous slope form can be attributed to deposition combining gravitational and slope wash processes. Weathering processes and increased slope wash has removed the easily soluble matrix. Consequently, clastic debris originating from the terrace deposits has accumulated by rock falls and colluvial slides at the foot of the terrace slopes. Subsequently, it has been distributed downslope by slope wash processes leading to the concave form. Similar to their parent material, the resulting deposits consist of fine-grained matrix and coarse clasts. Therefore these forms are interpreted as colluvial slopes.

However, these colluvial slopes are covered by comparatively dense vegetation. Particularly the occurrence of large cacti point to a relatively old age of these slopes. In addition, at many places the colluvial slopes show a pronounced scarp of several meters height resulting from lateral floodplain activity (Fig. 77). Finally, gullies cut through the colluvial slopes indicating that the slopes have been subject to intense fluvial dissection (Fig. 78).

Altogether, the colluvial slopes are interpreted to be relict and presently degrading landforms. During their formation wetter climatic conditions are likely to have restricted linear fluvial activity within the slope areas and on the floodplain, and to have enhanced slope denudation by wash processes. Resulting in a smoothed slope morphology, these processes are typical for climates of increased humidity. The dissection of these colluvial slopes reflects a renewed change to more linear fluvial activity and erosion, possibly as the result of increasing aridity, which would thin out the vegetational cover and increase the concentration of runoff. However, the dissection of these colluvial slopes has likely not been a single event. While their distal part shows a pronounced scarp of remarkable lateral continuity, their linear dissection led to the deposition of small alluvial fans onto the floodplain. Therefore a phase of floodplain incision must have preceded the dissection and deposition of the fans.

PRELIMINARY RESULTS

Regarding their above-described morphological characteristics, the slopes of the depositional terraces have been and continue to be highly dynamic areas. Several features like clay coatings and earth pyramids, but also fluvial processes like gullying, undercutting and alluvial fan deposition proove badland formation to be an active component of the present morphodynamic landscape of the study area. Particularly for the small number of inhabitants of the Quebrada de Purmamarca, this implies an enormous risk potential. In addition, it certainly presents limitations for all kinds of agricultural land use and infrastructure. Aside from the present dynamics of the badland areas, the terrace slopes have recorded evidence for geomorphic change attributable to changing environmental conditions. In this context, relict colluvial slopes have been suggested to indicate relatively wet conditions, while landslides may reflect a phase of increased slope instability during or following a phase of intense fluvial incision, possibly linked to increased moisture.

4.4. SLOPE MORPHOLOGY

Within the entire study area overall slope morphology of mountains and mountain chains is characterized by two contrasting geomorphological zones (Fig. 79). The mountain tops are dominated by gentle, convex slopes of smooth morphology. Downslope, this zone passes over into steep, concave and dissected slopes, sometimes in their lower part covered by fluvial terraces and/or alluvial fans. This striking morphological difference is apparent beyond the study area within large parts of the Cordillera Oriental (SEGEMAR-ITGE 1998).

Fig. 79: Panoramic view of the Cerro del Cobre, viewing WNW. Note how slope morphology varies from the mountain top to the valley floor.

4.4.1. PERIGLACIAL MORPHOLOGY

A round, sometimes flat topography and a very smooth relief are the dominant morphological characteristics of the higher mountain chains in the study area. This is particularly true for chains of more than approximately 3,400 m.a.s.l., e.g. Cerro del Cobre (3,905 m.a.s.l.) east of the Quebrada de Estancia Grande, Cerro Azul Alto (4,376 m.a.s.l.) east of the Quebrada de Huachichocana (compare geomorphological map) or Abra de Lipán (4,105 m.a.s.l.). Due to the enormous relief and extent of these slope areas, detailed field observation has been virtually impossible for the most part. Therefore, with the exception of observations made in vicinity of the Abra de Lipán, the following observations and results are mainly based on remote sensing data.

The highest parts of the study area are located in the upper part of the Quebrada de Estancia Grande (up to 5,036 m.a.s.l.) and west of the Quebrada de Sepulturas around Cerro Morado (up to 4,577 m.a.s.l.). The morphology in both areas is characterized by steep, straight to concave slopes. From satellite data and aerial photography slope angles have been stereoscopically estimated to 25° - 40°, locally up to 50°. Mountain tops appear as relatively pronounced crests and ridges above approximately 4,400 - 4,500 m.a.s.l.. These ridges do not have a sharp, angular morphology. Instead the ridge tops are smooth, possibly due to layers of debris covering the surface. The transition to the extended slope areas is characterized by outcropping rock spurs or cliffs not more than a few tens of meters in size. Immediately below the crests and ridges these spurs or cliffs are particularly frequent (Fig. 80). In all cases, these rock spurs are located in positions with east, southeast and south aspects. The frequency of these features decreases downslope, and below approximately 4,400 m.a.s.l. they have not been observed anywhere in study area. However, they are associated with and surrounded by slopes of various angles up to 50°. They exhibit a very smooth surface morphology, their coloration seems almost blurred, different colors grade into each other. This surface texture has only been observed at elevations above 4,400 m.a.s.l. while below color shades are relatively uniform, with sharp boundaries to each other. This is supposed to result from the gradual mixing of debris of different lithological characteristics during the downslope transport.

Nevertheless, the smooth and gentle appearance of slope morphology extends to elevations of much lower than 4,400 m.a.s.l., but thereby passes into increasingly convex slope profiles. On the available remote sensing data it could locally be traced down to elevations of approximately 3,400 m.a.s.l., e.g. at Cerro del Cobre, Cerro Estancia Grande, Cerro Lipán and Cerro Azul Alto. It has been observed in areas of different lithologies and aspects, possible slightly pronounced in northerly aspect, e.g. Cerro de Estancia Grande or Cerro Azul Alto.

The German word "Glatthang" (*German: "glatt" = "smooth", "hang" = "slope"*) has traditionally been used for a type of high mountain morphology showing the above described characteristics (e.g. HAGEDORN 1970, STINGL AND GARLEFF 1983). Therefore, large areas of the study area can be associated with processes of glatthang formation resulting in a typical smooth and regularized glatthangrelief. The smoothing of slope morphology is the result of the combined action of intense frost cracking, frost creep and gelifluction. Within a broad altitudinal zone, glatthang formation is typically active with different intensities.

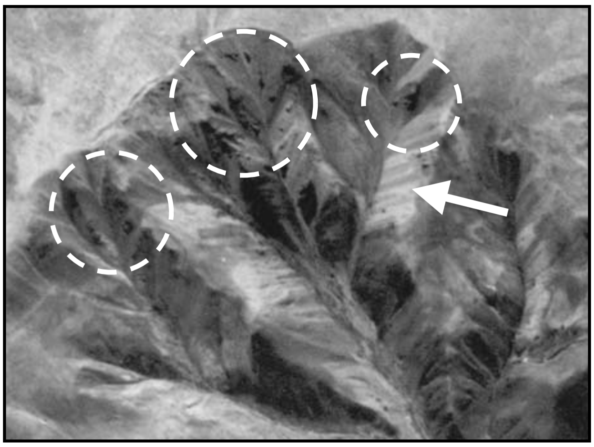


Fig. 80: Glatthang relief between 4,100 and 4,500 m.a.s.l. at Cerro Morado; note blurry and smooth slope appearance (arrows) interspersed with rock cliffs in circles (CORONA imagery, image width ~3,300 m).

Fig. 81: Gelifluction lobes above 4,300 m.a.s.l. in the upper Quebrada de Lipán; arrow indicates downslope direction (GASATACAMA aerial photography, image width ~700 m).

On the slopes at an altitude of 4,300 – 4,400 m.a.s.l. large lobe-shaped features have been observed in the northern study area, particularly in the Quebradas de Sepulturas and Lipán (Fig. 81). The lobes are 50 to 100 meters wide and up to several hundred meters long. They extend downslope from crests and ridges. Considering their lobe-shaped form, these features may be interpreted as gelifluction lobes. The slow, downslope transport of a moisture saturated mass requires the presence frozen ground or a considerable moisture content in the soil. Due to the semi-arid climatic characteristics of the study area, the saturation of the soil is attributed to freezing and thawing cycles of the uppermost soil horizon above frozen ground. Therefore active gelifluction would indicate the lowermost existence of frozen ground or discontinuous permafrost in the study area.

At least within the uppermost soil horizons, permafrost has not been observed within the study area at elevations below 4,100 m.a.s.l.. Instead, roadcuts several years old do not show any cryogenic disturbance, a fact which points to the absence of permanently or at least semi-permanently frozen ground. Only in the highest part of the Abra de Lipán at almost 4,200 m.a.s.l., water has been observed to escape from a roadcut, creating small isolated spots of enhanced vegetation. This excess water may be attributed to the damming effect of a frozen layer within the ground. In combination with the existence of active gelifluction at elevations above 4,300 m.a.s.l., these places could possibly be interpreted as indicators for discontinuous permafrost.

Fig. 82: Typical slope of dominating frost creep and gelifluction ("Glatthangrelief") close to the Abra de Lipán, 4,100 m.a.s.l.. Note zones of different color shades; agricultural use of the valley floor indicates limited activity of present periglacial processes.

Fig. 83: Glatthangrelief at ~4,000 m.a.s.l. close to the Abra de Lipan (aspect towards the S). Note relatively dense vegetational cover and inactive road cut.

Similar to the higher parts of the study area described above, the slopes in the vicinity east of the Abra de Lipán show morphological characteristic of glatthang relief (Fig. 82 and 83). Slopes are covered entirely with coarse and very angular clast debris resulting in a slope smooth appearance. Some bunch grasses populate these slopes. Slight differences in slope coloration can be noted, but by no means as evident as in elevations above 4,300 m.a.s.l.. These differences have been attributed to variations in vegetation cover and clast lithology, as no patina was found to coat the clasts. However, unlike the desert pavement typical for the surfaces of the depositional terraces, a remarkable amount of fine-grained material covers the surface in between larger clasts and bunch grasses. The slope profile is mostly straight to slightly convex, while slope angles vary between essentially no inclination to up to 40°. Again, this type of smooth morphology has been observed on slopes of very different lithology. While it has been found in positions of any aspect, it seems to have developed in a slightly more pronounced way on slopes roughly facing north.

While gelifluction describes the downslope movement of saturated debris sheets (*German: "Hangschuttdecken"*) above frozen ground, frost creep refers to the movement of debris by contraction and expansion due to freezing and thawing. Frost creep is a periglacial process not necessarily depending on the existence of permafrost. However, it is inferred to be of relatively low importance in the study area below 4,100 – 4,200 m.a.s.l.. As mentioned above, none of the roadcuts below these elevations have shown observable indications for active particle transport. In addition, up to approximately 4,000 m.a.s.l., small agricultural fields at the foot of slopes with glatthang morphology do not seem to be subject to accumulation or disturbance by periglacial processes (Fig. 82). Therefore it is concluded that even though the glatthangrelief is characteristic of the overall slope morphology down to elevations of around 3,400 m.a.s.l., periglacial processes are presently reduced to a minimum amount of activity at below 4,100 – 4,200 m.a.s.l..

At 3,500 m.a.s.l. a further roadcut gives insight to vertical differentiation within the slope (Fig. 84). The base is formed by grayish to black bare rock of Precambrian schists. The overlying horizon of brownish to grayish color consists of poorly sorted angular clasts, mainly schists and a few quartzites. The average clast size ranges between five and ten centimeters and the horizon is between one and three meters thick, with minimal lateral variation in thickness. By contrast, the topmost layer is up to six meters thick in its central part. Laterally, its thickness decreases to a few decimeters. This layer of grayish color consists of very poorly sorted angular clasts of up to 50 centimeters in diameter. Most clasts are of quartzitic lithology. The average clast size is approximately 15 centimeters. The entire layer is remarkably rich in fine grain sizes, while its lower part is dominated by coarse clasts of 30 to 50 centimeters. The sedimentological and morphological characteristics of this roadcut lead to the conclusion that a pre-existing topography has been smoothened in at least two steps indicated by the different layers in the roadcut. The lower layer is interpreted to be the product of intense frost creep and glatthang formation, as most of its material originates from the underlying schists. Disintegration of the bare rock by frost cracking leads to a horizon of coarse, angular clasts with accumulated finer grain sizes. This horizon, inherent to glatthang formation, has been reported to reach several meters of thickness (STINGL AND GARLEFF 1983). On active glatthang slopes this horizon is overlain by a thin cover of predominantly fine grain sizes, which corresponds to debris sheets moving downslope as a gelifluction mass. Due to its high amount of fine grain sizes, the upper layer may indicate the cumulative effect of gelifluction processes, filling up existing depressions and resulting in the evident topographical smoothing. Presently the roadcut does not show any signs of disturbance, a fact that emphasises the inactivity of glatthang formation and frost creep at this elevation.

Fig. 84: Colluvial material of relict frost creep in the Quebrada de Potrerillos, 3,500 m.a.s.l. Note how the material has covered and smoothened a pre-existing topography.

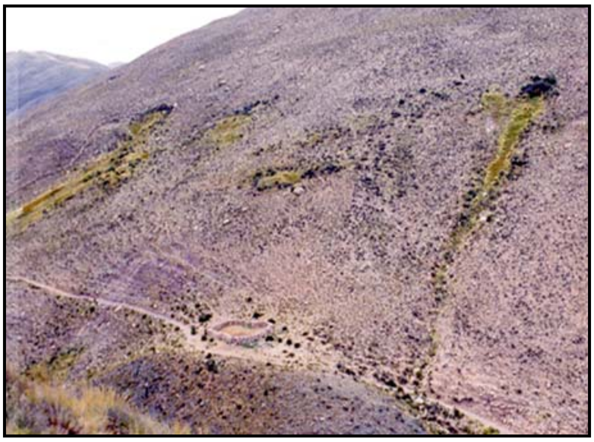


Fig. 85: Springs at ~3,500 m.a.s.l. in the Quebrada de Sepulturas, possibly related to a cover of relict frost debris sheets.

At an elevation of approximately 3,500 m.a.s.l. several springs have been observed in the Quebrada de Sepulturas (Fig. 85). These places show up as spots of enhanced vegetation, no more than several meters wide but extending 40 to 50 meters downslope. All of them have been observed aligned at similar elevation. Their position coincides with a pronounced change in slope angle. Above, slope angle is approximately 20° to 30° and below 40° to 50°. The slopes are covered by very few grasses. Otherwise most of the surface is covered with a layer of uniformly colored, relatively coarse debris, and no outcrops of bare rock are visible.

Judging from their setting and the associated morphological characteristics, it seems likely that the springs correspond to the lower end of the debris layer covering the slope above this elevation. Due to the change in slope angle, which is very likely the result of fluvial erosion, the debris layer crops out allowing the water which has collected inside to escape. Therefore the springs are indicative of the fact that slopes are covered by extended debris layers, at least above an elevation of 3,500 m.a.s.l..

Fig. 86: Mats of vegetation at Potrero Grande, 3,350 m.a.s.l.; Note the undisturbed road cut implying a small amount of present slope activity.

At an elevation of 3,350 m.a.s.l. several lobes of enhanced vegetational cover have been observed (Fig. 86). Each of these features has a diameter of five to 15 meters. From their morphological appearance they may be interpreted as typical products of gelifluction. Within this altitudinal belt, no similar features have been observed anywhere in the study area. In addition, despite its apparent age of several years, the roadcut just below again does not show any signs of disturbance. Therefore, these lobe-like features are interpreted as mats of preferential growth of vegetation due to a favorable northward aspect.

On top of the terrace surfaces, particularly the older generations T-1 and T-2 at Potrero Grande between elevations of 3,150 – 3,300 m, the V-shaped drainage valleys show a clear asymmetry of their slopes (Fig. 87). The valleys are between 100 and 200 meters wide and have cut into the coarse-clastic fanglomeratic terrace deposits. The valley orientation varies between WNW-ESE to NW-SE. In all cases the southern slope with its northward aspect has been found to show significantly lower slope angles than the northern slope, sometimes with differences of 15° - 20°. Absolute inclinations range from 20° - 25°, while those of the northern slope range from 30° to 45°. Furthermore the asymmetry is reflected by the relative distance from the valley shoulder to the drainage channel, deviating by up to 50 %.

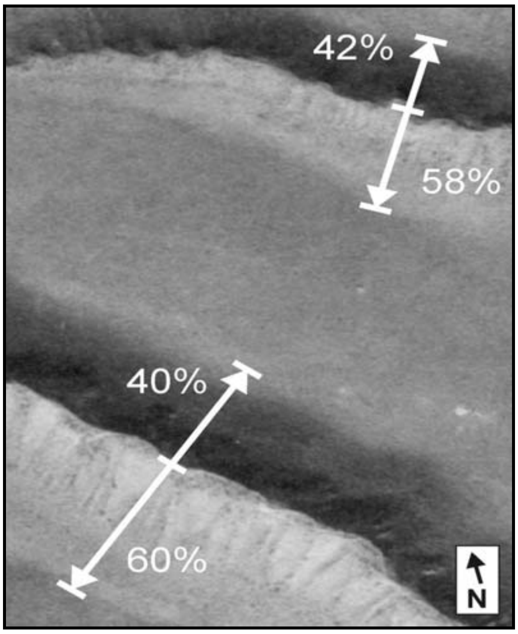


Fig. 87: Asymmetric slopes of drainage channels on terraces at Potrero Grande, interpreted to reflect phases of intensified frost creep and selective periglacial denudation due to differences in aspect (width of image ~400 m).

While this asymmetry is most evident on top of the older terraces, the drainage channels on the surface of terrace T-3 do not show any indication of asymmetry. In any case, periglacial processes are thought to be responsible for selective denudation, which has caused the asymmetry. Slopes with a northward aspect are likely to have experienced more frequent and intensified freeze-thaw cycles. Thus, pronounced frost creep and gelifluction should have resulted in an increased denudational efficiency on these slopes. Therefore the asymmetry of the valley cross profile may very well be attributed to periglacial reshaping of the terraces T-1 and T-2.

Finally, frost cracking is a very commonly observed process in the entire study area. The vulnerability of the rocks varies, depending on lithological characteristics. Fine-grained clasts of the Precambrian Puncoviscana Formation and the Ordovician Santa Victoria Groups are easily parted into thin fragments along foliation planes while quartzitic and sandstone clasts from the Mesón Group seem to disintegrate much slower (Fig. 88 and 89).

Fig. 88 and 89: Frost cracked rocks of Cambrian quartzite (left) and Precambrian schists and slates. Note multiple cracking along foliation- and/or bedding planes in the Precambrian schist.

In addition to rock lithology, the intensity of frost cracking depends on the frequency of freeze-thaw cycles. As a direct result of differences in elevation, temperatures in the upper study area drop below 0° C almost daily, while in the lower study area frost only occurs during up to 150 days (2.4.3.). Consequently, these climatic conditions imply a difference in production rates of frost debris with highest rates on the higher slopes and mountain chains.

From the above-described observations of periglacial morphology, several preliminary conclusions can be drawn. Landforms indicating periglacial conditions have been found to be characteristic for all higher mountain chains in the study area. The present altitudinal transition to a zone of at least discontinuous permafrost can be assumed to lie at an average of 4,200 m.a.s.l. in the upper study area. Above this elevation active gelifluction has been observed. Presently active slope smoothing is inferred from the blurred coloration of many slopes above approximately 4,400 m.a.s.l. and interpreted as evidence for active glatthang formation. In the highest parts of the study area, glatthang relief is increasingly interspersed with outcrops of rock spurs. While surrounded by slopes typical for glatthang formation, these features might be interpreted as remnants of a more pronounced previous topography, possibly being the result of periglacial processes other than glatthang formation.
Even though no active glatthang formation is presently evident below 4,400 m.a.s.l., widespread relict glatthang morphology extends down to elevations of approximately 3,400 m.a.s.l.. In addition, a variety of other relict periglacial features have been observed in elevations between 3,100 and 4,200 m.a.s.l., indicating a severe drop of the periglacial belt in the past. Apparently, this shift of geomorphic and very likely climatic regimes predates or parallels the accumulation of the last terrace generation T-3, as the older terraces exhibit

convincing evidence for periglacial reshaping. Summing up, a severe drop of periglacial conditions would have enormous effects on the production and transport rates of frost debris, a fact that is particularly interesting with regard to the extraordinary size of the depositional terraces described above.

4.4.2. GLACIAL MORPHOLOGY

In contrast to the periglacial conditions described above, the presence of glacial conditions depends on temperature *and* precipitation. Therefore the present snowline in the semi-arid study area is expected to be located at elevations significantly higher than the periglacial zone. This fact alone constituted a great limitation to field observation of glacial morphology. Nevertheless, some important conclusions can be drawn from the observations based on the analysis of remote sensing data.

As mentioned above, even the highest parts of the study area exhibit characteristics typical of periglacial morphology. The presence of glacial features within the Quebrada de Purmamarca today can therefore be denied. Nevertheless, some authors mention glacial related features such as cirques and moraines from the Sierra Alta, a range of similar elevation in the direct vicinity north of the study area (SEGEMAR-ITGE 1998). Even though the highest ranges in the upper part of the Quebrada de Estancia Grande (up to 5,036 m.a.s.l.) and west of the Quebrada de Sepulturas around Cerro Morado (up to 4,577 m.a.s.l.) do show a morphology dominated by crests, ridges and concave slope profiles, further evidence for glacial processes is absent.

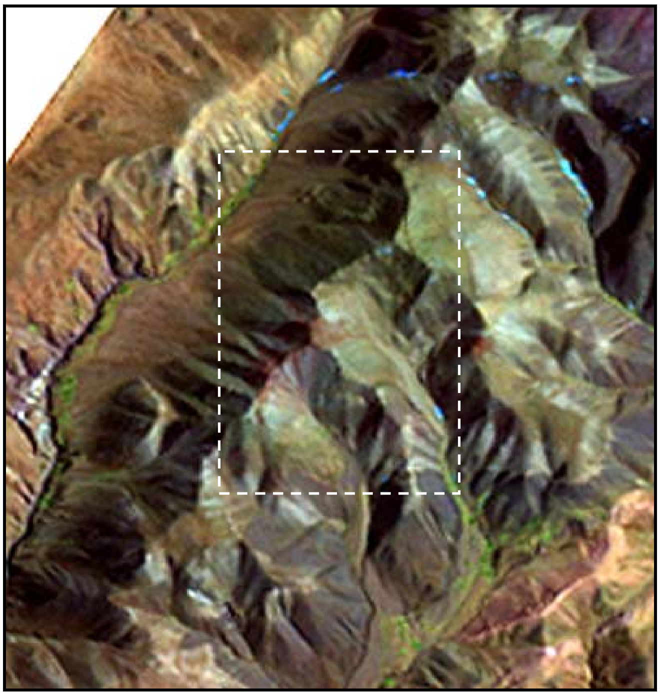


Fig. 90: LANDSAT TM 5 image from the northern Quebrada de Estancia Grande (up to 5,036 m.a.s.l.). Note the markedly sharp ridges (width of image 5,500 m).

Fig. 91: Enlarged detail of Fig. 90 from high-resolution CORONA satellite data (image width ~1,800 m). Note outcrops of rock spurs below ridges and the smooth morphology of ridges.

While the overall morphological appearance of these areas seems relatively sharp and angular on Landsat TM 5 data, a comparison with high-resolution Corona satellite data and aerial photography reveals that ridges and crests are evidently smooth and regular (Fig. 90 and 91). This example shows that geomorphological mapping of regions not accessible from Landsat data alone carries certain risks and is always prone to misinterpretation. In addition, from the downslope profiles of valleys within the highest parts of the study area neither a cirque depression nor a cirque threshold could be inferred. Due to the absence of these forms, the existence of cirques in the study area can be ruled out.

Finally, depositional glacial forms have not been identified anywhere in the study area, either. Exclusively in the eastern Quebrada de Sepulturas, at an elevation of approximately 3,700 m.a.s.l., an elongated ridge-like feature 700 meters long has been detected from aerial photography (Fig. 92). It has been estimated to be 40 meters wide and approximately 15 meters high. The surface of this feature appears to be relatively smooth. It is located upstream of the confluence of two valleys along the southern foot of a hillslope and stands 40 – 50 meters above the present valley floor. Upstream the valley continues into the area dominated by glatthang morphology up to the Cerro Morado, with 4,577 m.a.s.l. being the second highest peak in the study area.

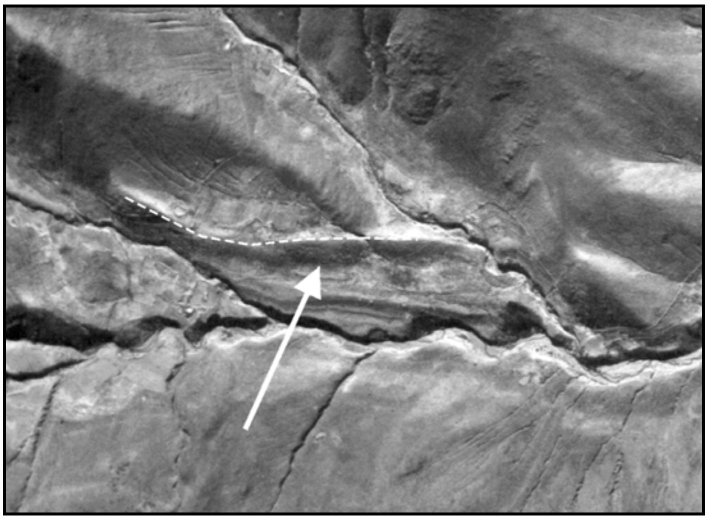


Fig. 92: Feature in the Quebrada de Sepulturas (~3700 m) showing simil-arity to a lateral moraine (GasAtacama air photography, width of image ~1,600 m)

Judging from its morphological characteristics, the feature might be regarded as a lateral moraine. However, no comparable forms have been found in any other place. No further lateral moraines or even terminal moraines correspond to the feature.
Therefore it seems reasonable to attribute the formation of this form to processes other than glacial deposition. From its morphological context it might be interpreted as a high and reshaped remnant of a fluvial terrace. Alternatively, its formation may have been the product of accumulating debris at the foot of a snowfield, which would have persisted for a longer period of time due to the southern slope aspect. In any case the unavailibity of field data bears certain limitations for the definite interpretation of this feature.

Summing up, available data seem to imply that the study area has not experienced glaciation at any time in the younger landscape history. Whether the area has been subject to glaciation during much earlier stages of landscape evolution cannot be decided.
While the annual amounts of precipitation decreases from east to west, the modern snowline in NW-Argentina increases towards the west and should therefore be located at around 6,000 m.a.s.l. in the semi-arid to arid study area (FOX AND STRECKER 1991, HASELTON ET AL. 2002). However, glacial conditions do not seem to have prevailed at any point in the study area. This corresponds to the fact that no active glaciers have been observed or mapped.
ZIPPRICH (1998) has mapped several stages of moraines in the Sierra de Santa Victoria, the lowest (and oldest) of which is located at 3,850 m. Corresponding Pleistocene snowlines have never decreased to below 4,300 – 4,650 m. However, the Sierra de Santa Victoria is located much further east than the study area and receives precipitation of slightly more than 300 mm. This suggests a significantly higher snowline for the study area due to lower precipitation. For all of NW-Argentina FOX AND STRECKER (1991) and HASELTON ET AL. (2002) assume a Pleistocene drop of snowline of 300 – 900 m. GARLEFF AND STINGL (1985) assume a maximum drop of 400 m. Therefore, combining the geomorphological results discussed above with available regional literature, the study area is unlikely to have experienced any form of glaciation during the Late Pleistocene.

4.4.3. SLOPE DISSECTION

While the higher parts of the mountain chains in the study area are characterized by periglacial morphology, the next lower slope segment has been observed to be subject to intense forms of fluvial dissection. Generally, three different types of dissection have been evident from field observations.

The first type of dissection refers to well developed fluvial channels, sometimes with inclinations of up to 30° depending on the local slope angle. Usually these channels show typical signs of a drainage network evolution, i.e. the branching into minor channels and a well-definable drainage catchement (Fig. 93). The main channel is between several hundred meters up to a few kilometers long and starts in the topmost, convex parts of the mountain chains. It can be incised into the slope several tens of meters. This type of dissection has been observed in all parts of the study area, reaching up to a maximum elevation of nearly 4,000 m.a.s.l.. In most cases the drainage channels have a direct connection to the present valley floor, but occasionally they grade into relatively large alluvial fans downslope. However, regarding their size and well-developed drainage pattern, these channels are interpreted as an integrative part of the drainage system of considerable age, very likely predating the accumulation of depositional terraces as well as the intense periglacial processes responsible for the slope morphology in the higher study area.

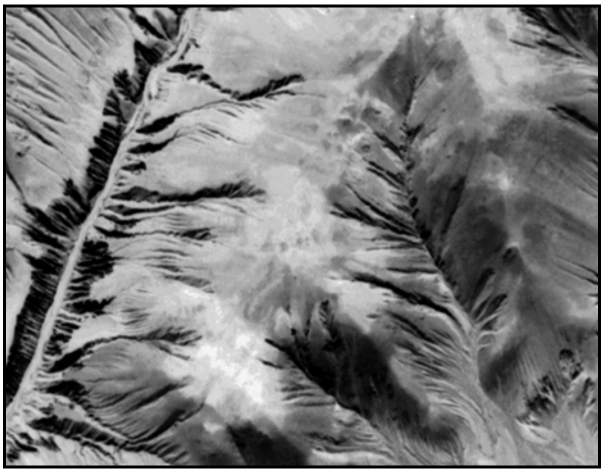


Fig. 93: Slope dissection by well-developed drainage channels in the Quebrada de Estancia Grande. Note branching into tributaries (CORONA imagery, width of image ~2,900 m).

The second type of dissected slope is characterized by relatively deep, almost parallel and V-shaped drainage channels (Fig. 94). The steep and very straight channels have incised up to 20 meters into the slope and usually have an inclination of 20° - 40°. The spacing between the individual channels ranges from several tens to up to hundreds of meters, dissecting the slope into a series of parallel ridges and channels with a spacing of a few tens of meters. Their initial part corresponds in all cases to the transition of a concave to a convex slope profile. Sometimes they reach up to 400 meters down the slope, where they run out onto a terrace surface or form a presently active alluvial fan (Fig. 94).

Fig. 94: Dissected slope in the Quebrada de Lipán running out in alluvial fans on the T-3 terrace surface. Note debris cover on slope areas between the channels (3,300 – 3,500 m.a.s.l.)

Fig. 95: Dissected slope in the Quebrada de Estancia Grande between ~3,200 - 3,500 m.a.s.l.. Note the removal of debris cover leading to the exposure of bare rock.

This type of dissection has been observed independent of aspect on slopes of varying lithologies, but predominantly on quartzitic slopes in the upper study area, commonly at elevations between 3,000 and 3,800 m.a.s.l.. Even though the channels have clearly incised into bedrock, the ridges and slope segments in between the channels are apparently covered with a thin layer of debris and some vegetation. Therefore the dissection of the slopes can be inferred to postdate the formation of the debris sheets covering the slopes.

Regarding its morphological characteristics, the third type of dissected slope is very similar to the second type described above. Dissection occurs in a series of almost parallel, V-shaped drainage channels of varying depth and size on slopes between approximately 20° and 40° (Fig. 95). While some channels extend downslope as far as 500 meters, most of them are not longer than 150 to 200 meters. Channel width varies between a few meters and several tens of meters. Corresponding to their length, channel incision into the slope has been found to be between five and more than 30 meters. Because of their varying length some channels have been observed to initiate in relatively high and convex slope areas, but similar to the dissection type described above, channel heads mark the transition between convex and concave slope segments.

The spacing of channels is very variable; smaller channels can show a spacing of several meters, while the larger channels are separated by non-dissected slope segments several tens of meters wide. Contrary to the dissection type described above, this dissection type shows a slight branching tendency. Some of the larger channels have developed tributaries. In addition, up to a certain elevation, the dissection seems to have denuded and removed the entire sheet of slope debris ("Hangschutt"), areally exposing the bare rock. Above, where the slope debris still covers the slopes, some of the channels continue as narrow and shallow gullies without showing the typical V-shaped form of the dissection in the lower part. However, downslope most of the larger channels run out into active alluvial fans while the smaller channels have apparently been cut/blocked by the deposition of these fans (Fig. 95). Where no alluvial fans exists, the channels have a direct connection to the floodplain. Independent of aspect, this type of dissection has been observed exclusively on Precambrian schists and phyllites, predominantly in the upper study area, particularly between approximately 2,700 – 3,500 m.a.s.l.. Therefore it is assumed to benefit from lithological characteristics of these Precambrian rocks.

As mentioned above, the second and third type of dissection is characteristics of the upper study area, particularly between 2,700 – 3,800 m.a.s.l.. However, they show a number of morphological differences, particularly with regard to the erosion and denudation of slope debris. Apparently these differences are caused by lithological differences, as each of the dissection types has been observed on slopes of different rock lithology. The fact that rock lithology is a major control on fluvial slope dissection is emphasized by a further observation.

A fourth type of dissection has been found to be restricted to the small outcrop areas of Ordovician slates at La Ciénaga and in the Quebradas de Chalala and Coqueña (Fig. 96). The slates have very fine-grained lithology and are rich in clay, which makes them particularly vulnerable to weathering and fluvial erosion. In addition these rocks constitute relatively poor locations for vegetation, which increases the potential for denudational and erosional processes. The morphology of this dissection type is characterized by a high number of regularly distributed V-shaped channels and relatively gentle, smooth slopes, separated by sharp ridges. Slope angles are constantly between 30° and 40°. Judging from these characteristics these areas can be classified as badland; badland formation in clay-rich rocks is indeed a commonly observed phenomenon in drylands (CAMPBELL 1997).

Fig. 96: Overall impression of badlands in Ordovician slates at La Ciénaga.

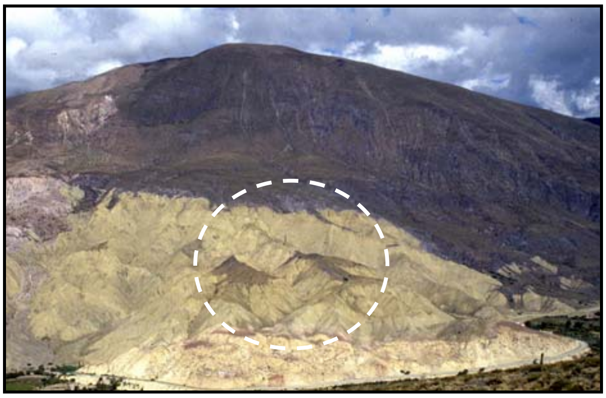


Fig. 97: Panoramic view of badlands in Ordovician slates at La Ciénaga (direction towards the W); note remnants of debris sheets (indicated by circle).

Due to the intensity of fluvial erosion, vegetation as well as rock debris are entirely absent within these badland areas, exposing the bare rock and causing the intense yellowish coloration of the badlands. Upslope the badland area at La Ciénaga is bounded by outcropping Precambrian schists.

Nevertheless, at two places of the badlands at La Ciénaga, flat and smooth surfaces of a brownish to greyish color have been preserved - their slope angle has been estimated to 25°. These isolated surface remnants are approximately 100 meters long and wide and have been found at an elevation of 2,750 – 2,800 m.a.s.l. (Fig. 97). Their color points to the predominance of schists in these areas. These can only have reached their current isolated position by transport following a slope eroded since then. Therefore these remnants are interpreted as the last remaining morphological evidence of a pre-existing slope, which has been removed by fluvial dissection that created the presently active badland morphology.

Finally, another remarkable type of dissection has to be mentioned, which is particularly evident in the upper study area. At several locations, slopes are dissected by individual, sometimes even single drainage channels. Contrary to the above-described types of dissection, these channels have not been found in close association to adjacent channels. In addition they differ in morphological appearance. The dissection types described above have a marked influence on slope morphology; they have either cut relatively deep and pronounced V-shaped drainage channels by linear erosion or have extensively dissected the slope surface (Fig. 94 and 95). However, this type of slope dissection is limited to clearly defined drainage lines of maximum depths around ten to 15 meters with steep to almost vertical channel walls. The length of these drainage lines varies ranges from a few hundred to up to some kilometers. Their upslope head is not restricted to a certain slope profile. Some channel heads are found in the upper, convex slope segments while others initiate in the lower, concave slope segment or at the transition between both segments. From their morphological appearance these channels could be classified as gullies due to their narrow and steep channel forms (CAMPBELL 1997). These gullies have been observed predominantly in the higher study area between approximately 3,000 – 4,000 m.a.s.l. in various lithological settings. Most of these gullies have incised several meters into bedrock. Most probably the necessary sediment load for fluvial incision originates from weathering of the gully walls or has its source in the debris-covered higher slope areas, where slopewash and creep processes slowly contribute material to the gullies. In all cases the gullies are adjusted to the present valley floor, sometimes cutting through terraces or alluvial fans. Upslope, the channels show a marked transition from the deeper and incised channel part to a relatively narrow and shallow channel part of a few meters in width. This abrupt change is called headcut and is typical for the headward propagation of gullies, which indicates the present activity and ongoing extension of the gully (THORNES 1994).

In the lower parts of the gully, where slope gradient decreases, levees of accumulated debris have been observed. Mostly these levees are not higher than one to two meters and have a lateral extent of several tens of meters. Nevertheless, they show that the gullies serve as transportation lines for debris-flows, even though they are the result of fluvial erosion and incision. This can be interpreted as yet another evidence for the high present activity of the gullies. In addition it points to the considerable amount of debris which is presently being evacuated and transferred from the higher slope areas into the floodplains and the lower reaches of the study area.

Fig. 98: Gullies in the upper part of the Quebrada de Potrerillos at approximately 3,900 m.a.s.l.; carved by water, the gully serves as transport line for debris-flows, too, as indicated by the levees in the lower right corner.

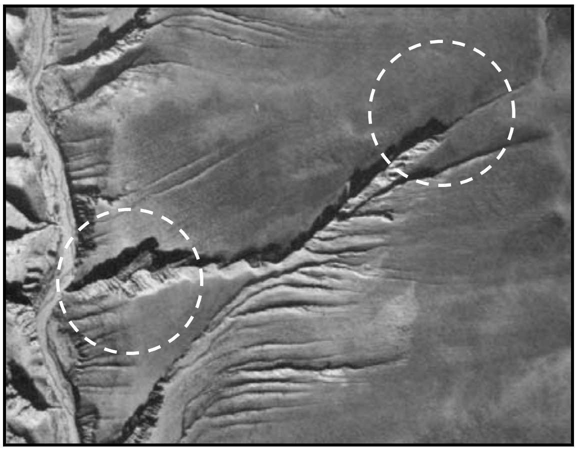


Fig. 99: Gully-like drainage line in the Quebrada de Lipán, approximate length is 800 m; note the transition of the straight, single channel into a narrow rill at headcut (right circle). Dissection widens where gully erodes terrace deposits (left circle) (GASATACAMA aerial photography, width of image ~1,000 m).

Summing up the above-described characteristic types and forms of slope dissection, several conclusions can be drawn. Below about 4,000 m.a.s.l. various forms of dissection and fluvial erosion are apparent. Some of the active dissection occurs through extensive and well-developed fluvial drainage channels, which are supposed to predate most periglacial slope processes as well as terrace accumulation and erosion. Significant slope areas are subject to active dissection by relatively small and narrowly spaced drainage channels. The channels initiate at the transition between the upper, convex slope segment and the lower, concave slope segment, very likely due to the acceleration and concentration of sheetflow

(THORNES 1994). They are particularly well-developed where slope angle is highest. Nevertheless, the channels show a clear V-shape and can therefore not be classified as gullies.

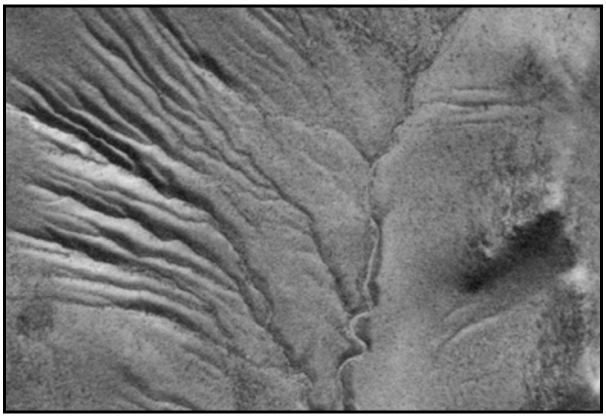


Fig. 100: Dissection of slopes of Cambrian quartzite in the Quebrada de Lipán; note the apparent adjustment of channels to the drainage network on the terrace surface (GASATACAMA air photography, width of image ~600 m).

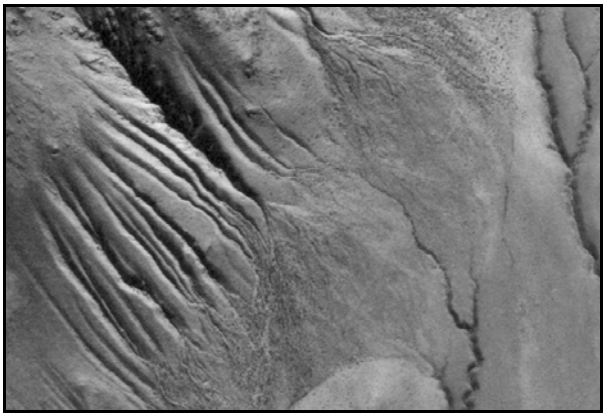


Fig. 101: Dissection of slopes of Cambrian quartzite in the Quebrada de Lipán; note the formation of alluvial fans on top of the terrace surface (GASATACAMA air photography, width of image ~900 m).

Depending on their location, some channels have adjusted to the terrace surface of T-3 (Fig. 100 and 101) or to the present valley floor, suggesting that their development postdated the accumulation of the deposition of the terrace T-3. Considering the active alluvial fans or connection to the existing drainage network, the dissection is inferred to be presently active in most cases. Even though they are actively dissecting the slopes, gullies have been observed adjusted to the valley floor in all cases. Similar to the gullies responsible for the dissection of the terrace surface of the T-3 terraces, these gullies are interpreted as relatively young and recent features, as indicated by their channel morphology.

With regard to the average elevations of the dissection some important conclusions can be drawn. While in the lower study area up to approximately 2,700 m.a.s.l. the dissection seems to be restricted to fluvial erosion within the well-developed drainage system, various types of dissection have been observed between 3,000 and 4,000 m.a.s.l.. While the overall smooth and gentle slope morphology above 3,400 m.a.s.l. had been inferred to be the product of past periglacial processes, the same slopes are now evidently being subject to intense dissection and erosion.

In some cases the dissection has initiated at transitions of the slope gradient and has adjusted to a pre-existing topography, e.g. the terrace surface. Very often, this type of dissection extends over larger slope areas, channels are V-shaped and have even developed catchments large enough to deposit extended alluvial fans. Therefore this type of dissection is considered to predate the dissection by gullies. By contrast, the gullies seem to have expanded headwards and continue do so. They are interpreted as relatively young features.

Considering the various theories of gully formation (e.g. THORNES 1994), the possibility of anthropogenically induced formation has to be discussed. Besides climate change, an anthropogenic reduction of vegetation cover causes significantly lower infiltration rates, higher runoff and lower soil resistance, which results in incision and gullying.

Indeed, the pastures on the terrace surfaces and extended slopes of the study area are subject to a transhumance pasture system. Herds of goats, sheep and even cattle have been observed in all parts of the study area. CHOROLQUE (1998) reports the remarkable high number of more than 15,000 sheep and goats and 1,000 piece of cattle for the entire municipality of Purmamarca of 2000 km^2. Possibly due to the gentle topography, pasturing is most intense in the higher study area, even above 4,000 m.a.s.l. (Fig. 102) where the ecosystem is reported to be particularly vulnerable due to the sparse vegetational cover (RUTHSATZ 1977). With regard to the future development of the region, overgrazing has been mentioned to constitute a severe problem (CHOROLQUE 1998). Other well-documented examples of regional anthropogenically induced landscape change have been reported from the Sierra de Santa Victoria in NW-Argentina (KULEMEYER 1998, SCHÄBITZ 1999).

Fig. 102: Shed for animals, mostly goats and sheep, at an elevation of approximately 4,000 m.a.s.l. close to the Abra de Lipán.

Where dissecting slopes, gullies have been observed in all quebradas between elevations of 3,000 and 4,000 m.a.s.l.. In addition, gullies are very common on the terraces surfaces of terrace generation T-3 in elevations between 2,700 and 3,200 m.a.s.l.. In both settings they have been noted to reach extents of up to a kilometer. Particularly on the terraces surfaces they seem to grade into deeper and almost gorge-like fluvial channels, which finally cut the entire terrace down to a depth of several tens to more than one hundred meters. Considering this enormous amount of channel incision it seems very unlikely that the removal of vegetation by overgrazing has been the trigger for gully incision and channel development. Even though they do not reach quite the size of the gullies on the terrace surfaces, the gullies on the upper slopes are still considerably large, sometimes up to more than a kilometer. As mentioned above they are always adjusted to the valley floor and their development has been retrocedent in most cases. They have not been observed in the slope areas where grazing is most intense and can therefore not have been caused by grazing either. However, the intense usage of the higher slope areas as pasture grounds may have eventually accelerated the evolution of the gullies contributing to the apparent dynamic dissection in higher the study area.

4.4.4. SLOPE DEVELOPMENT AND WEATHERING PROCESSES

A very isolated and minor form of slope erosion has been observed in the Quebrada de Huachichocana at an elevation of approximately 3,400 m.a.s.l.. In the narrow gorge-like valley at least three caverns have been discovered. The largest of them (Fig. 104) consists of a single, half-opened room of 20 meters width and eight to ten meters height extending 15 meters into the mountain. The smaller caverns (Fig. 103) are less than ten meters wide and only one to two meters high. Their interior extent could not be determined.

At various places the overhanging walls have collapsed, which is particularly evident at the largest cavern. The ground in front of the cave is covered with very angular clasts of varying size between 30 centimeters and several meters, most probably the result of a minor rockfall. Within all caverns the ground is covered with andesitic sand and debris. The smaller caverns seem to be filled with sediment up to a much higher level than the large cavern (Fig. 103), possibly because the observed rockfall has protected the cavern from increased sediment input. While sediment fill in the smaller caverns forms a flat and horizontal level, a small cone of gravitational deposits with a marked sorting of clast sizes has formed in the larger cavern. Otherwise, a slightly brownish patina has been observed to cover parts of the walls and ceilings inside the caverns. Considering these observation the caverns presently seem to be subject to deconstruction by slow gravitational processes. The considerable sediment fill as well as the patina rule out any ongoing formation of the caverns.

Fig. 103: Small cavern in andesitic rock in the Quebrada de Huachichocana at an elevation of 3,400 m.a.s.l..

Fig. 104: Rock fall from cavern overhang in the Quebrada de Huachichocana at approximately 3,400 m.a.s.l.. Note fresh and angular appearance of the rocks.

Two additional observations concerning the setting of the caverns may point the processes involved in their formation. First of all, the existence of these caverns is restricted to the small area of outcropping andesite. In addition, they occur exclusively within the south-exposed valley walls. This implies that their formation can at least partly be attributed to the combination of lithological and microclimatic controls. Particularly in granular rocks like andesites evaporation of pore water and dew leads to an exfoliation of superficial rock flakes generating small pits called Tafoni. These tafoni grow bigger by positive feedback due to increased shadow and humidity. They may grow to considerable sizes to form abris. Alternatively, they can be considered a starting point for enhanced erosion and denudation by fluvial and mass wasting processes. A combination of these processes may have been responsible for the formation of the observed caverns.

Fig. 105: Tafoni (circles) in andesitic rock at 3,550 m.a.s.l. in the Quebrada de Huachichocana. Note ongoing disintegration of the rock along joints.

Aside from the caverns, several small tafoni-like holes have been discovered on the andesitic walls in direct neighborhood to the cavern (Fig. 105). Most of the circular to semi-circular pits measure five to ten centimeters in diameter while very few have grown to sizes of 15 to 20 centimeters. Corresponding to their size, they extend no deeper than five to ten centimeters into the rock. Similar to the caverns, the tafoni occur exclusively on south-facing andesitic surfaces. In this context, the same processes may have contributed to the formation of both the tafoni as well as the caverns.

However, the tafoni have been observed to be limited to small areas of the wall (Fig. 105). Most of the wall appears very angular and sharp, possibly due to enhanced physical weathering processes. As characteristic for volcanic rocks, the andesite seems to disintegrate along preexisting joints. At some places, these joints have been found to cut through individual tafoni. Concluding from these observations, a clear relict character can be assigned to the tafoni as well as the caverns.

Regarding the question for paleoenvironmental conditions responsible for the formation of caverns and tafoni, two conclusions can be drawn. First of all, the intensity of physical weathering as observed at present must have been significantly reduced. In addition, a certain amount of moisture must have been constantly available. Despite the favorable microclimatic position (south aspect = shadow), this seems not to be the case under the present semi arid conditions. Therefore climate during tafoni formation can be inferred to have been considerably wetter. The timing of the tafoni and cavern formation is hard to tie into a certain range. No geomorphological or topographic correlation to any other landform seems possible due to the isolated location of the caverns and tafoni within the narrow gorge. However, a minimum age of 10 ka BP can be assigned to the caverns as archeological investigations have proofed the caverns to have been inhabited by that time (KULEMEYER, J. A. 1998, KULEMEYER, J. A. ET AL. 1999). Considering the relatively large size of the caverns, a significantly older age is likely.

4.4.5. MASS WASTING PROCESSES

Aside from the above described periglacial and fluvial processes, mass wasting processes have been observed to be of some importance for slope morphology. In all cases the resulting landforms have been detected in the lowest slope segments in direct transition between slope and valley floor.

In the Quebrada de Purmamarca north west of La Ciénaga at approximately 2,800 m.a.s.l. a pronounced scarp of approximately 30 meters divides the slope into two segments (Fig. 106). The upper slope segment of 15° inclination gradually passes over into the higher slope areas of the Cerro Azul Alto. The lower segment of similar slope angle appears as an enormous block of rock. It extends over an area 900 meters long and 400 meters wide and is dissected by several smaller drainage channels. Judging from the geologic map, the entire, debris-covered slope consists of schists and phyllites of the Precambrian Puncoviscana Formation, a fact that is corroborated by the typical greyish color. However, in the lowest parts of this enormous block, several ten to 20 meter-sized outcrops of yellowish rocks have been noticed, all of them highly dissected. These rocks belong to the Ordovician Santa Victoria Group, which is characterized by very clay-rich slates. It has been mentioned above in the context of strong fluvial dissection and badland formation.

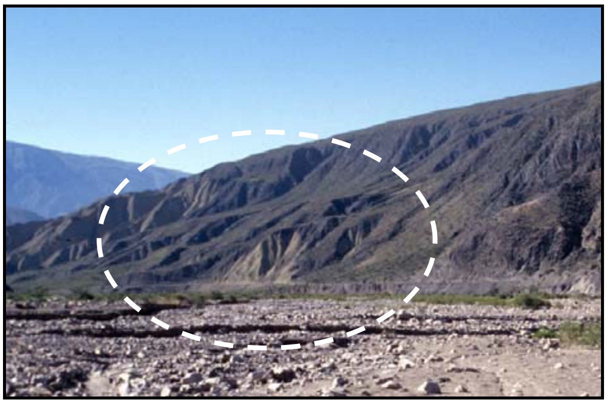


Fig. 106: Landslide at La Ciénaga, approximate elevation is 2,800 m.a.s.l.. Circle denotes the sliding mass.

Fig. 107: Landslide close to Patacal; approximate elevation 2,700 m.a.s.l.. Circle denotes sliding mass and black dots denote landslide scar.

Considering these morphological and lithological characteristics, the described block is interpreted as a relict landslide. Its relatively high age has been inferred from the dissection, which it has been subject to. However, the sliding event may have been triggered by a variety of factors. Its location at the convex ("prallhang") side of the floodplain may have caused intensified lateral erosion and undercutting leading to slope instability.

Alternatively, the slope instability may have been the result of the removal of great amounts of Ordovician slate by fluvial dissection of the easily erodable material. Finally, the entire block of overlying schists would have slid downslope, appreciating the fine-grained slates as ideal gliding planes. Theoretically, the slope could have been additionally weakened by an increased relief during a time of enhanced fluvial incision or by increased amounts of moisture infiltrating the slopes during a wet phase.

South of Patacal, opposite the confluence of the Quebrada del Cobre into the Quebrada de Purmamarca, the lowermost slope segment at 2,700 m.a.s.l. is divided into a large block of solid rock below a pronounced scarp approximately 50 meters heigh (Fig. 107). The block is not larger than 200 meters in length and width, but is estimated to be at least 50 meters high. It consists of Ordovician slates in its lower part and Cretaceous limestones in its upper part. The lower slope segment surrounding the block consists of slates, while the boundary to the overlying limestone corresponds remarkably well with the observed scarp. Comparing the original dip of the limestone, the block below shows a clear tilt towards the valley floor, which becomes obvious from a reddish marker horizon separating slates and limestone. Regarding the morphological and lithological characteristics, the described slope is interpreted as the product of a landslide. Particularly the participation of the clay-rich slates make this slope segment suspicious and may point to processes similar to the landslide described above. Again, the underlying slates may have been the subject to enhanced fluvial dissection, lateral undercutting and/or an increased moisture content, which consequently would have led to an increased slope instability being the precondition for each landslide. The setting opposite a relatively large alluvial fan coming out of the Quebrada del Cobre may point to pronounced diversion of the floodplain towards the slope causing intense undercutting.

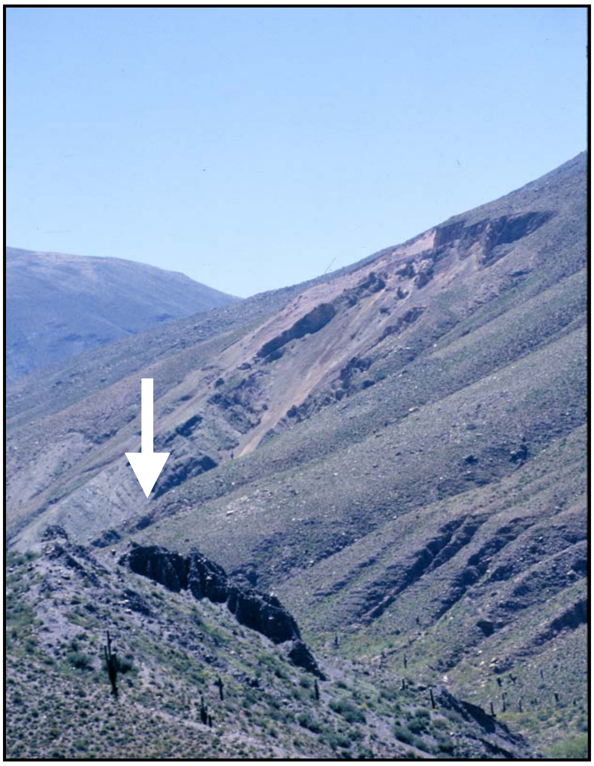


Fig. 108: Scar of rock slide (rock avalanche?) in the Quebrada de Sepulturas. Arrow depicts location of colluvial cones (Fig. 110).

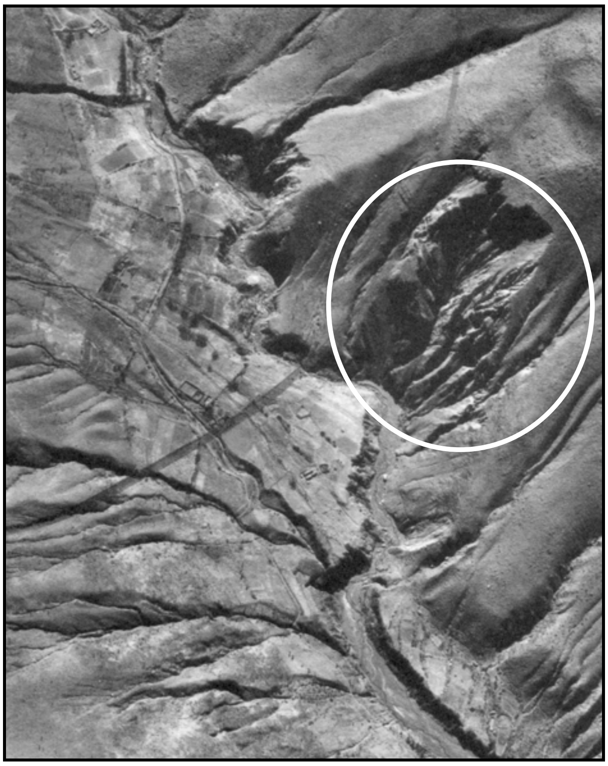


Fig. 109: Scar of rockslide in the Quebrada de Sepulturas. Note alluvial fan which has laterally diverted the drainage channel (GASATACAMA aerial photography, image width ~1,100 m).

Similar to the setting described above, at approximately 3,300 m.a.s.l. an alluvial fan in the Quebrada de Sepulturas has diverted the floodplain against the slopes to the east of the quebrada (Fig. 109). Opposite the fan, a 250 meter high, 100 meter wide and approximately 150 meter high scarp has been noted in the slope (Fig. 108). Judging from the geological map (SEGEMAR 1998) the entire slope consists of quartzitic sandstone and quartzites of the Cambrian Mesón Group. Apparently, the slope angle of approximately 40° corresponds to the dip of the sandstone and quartzite layers of the bare rock.

However, no deposits have been found in association with the scar, with the exception of two intercalated, small colluvial cones of 40 to 60 meters in height at the foot of the slope (Fig. 110). These colluvial cones have a slope angle of almost 50°. While the larger and thicker, slightly reddish one of these cones shows a vertical erosional scarp at its foot making it a clearly relict feature, the smaller, brownish to greyish one has been deposited aside the larger cone onto the present floodplain. In addition it partly covers an older lateral colluvial slope on which several shrubs grow. The colluvial deposits of both cones consist of a matrix-poor fanglomerate of angular clasts with an average clast size of five to ten centimeters while the amount of larger clasts is remarkable low. The dominant lithology of the clasts seems to be quartzitic, but several slates and schists have been observed as well.

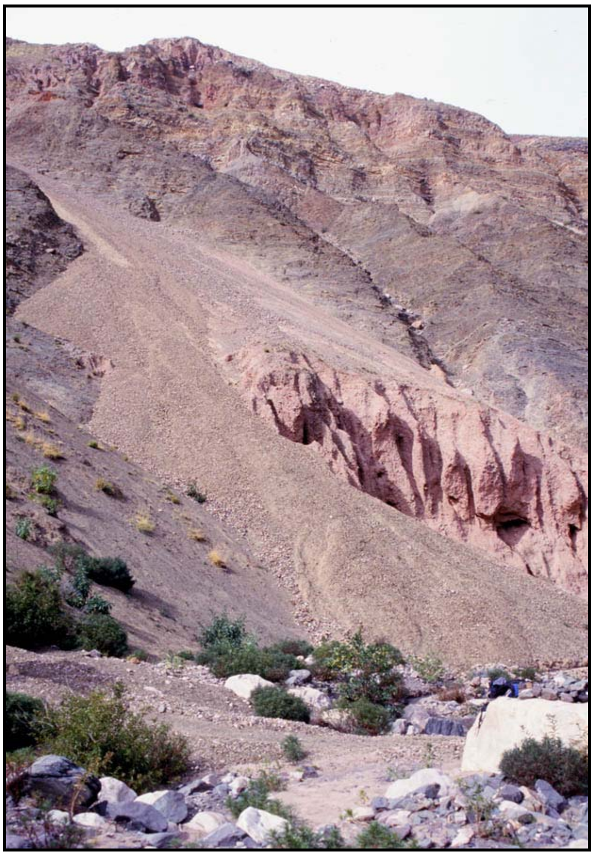


Fig. 110: Colluvial cones at the foot of rock slide scar in the Quebrada de Sepulturas at approximately 3,300 m.a.s.l..

Fig. 111: Scar of possible future landslide in the upper Quebrada de Potrerillos (GASATACAMA aerial photography, image width ~700 m).

Summing up the above-described characteristics, the observed scarp seems to give evidence of a mountain front collapse. Particularly the fact that slope angle matches the dip of local strata seems to indicate a setting prone to failure (Hermanns and Strecker 1999). The missing deposits of the collapse, however, do not allow conclusions about the type of processes. Very likely the dip of the strata has served as a sliding plane for large blocks of overlying rock, possibly even for the entire stratum. Judging from the elevation of the scarp, the vertical relief contrast has not been sufficient to generate a rock avalanche, as avalanches are estimated to form exclusively from failures of more than 400 meters (Hermanns and Strecker 1999). Therefore, this slope failure has been tentatively interpreted as a rockslide. Regarding its setting, lateral undercutting and intensified incision are very likely to have played a role in increasing the shear stress on the slope.
From the existence of the two colluvial cones at the foot of the scarp several conclusions can be drawn. First of all, a relatively high age can be inferred for the rockslide as it predates the formation of two colluvial cones. The formation of the two colluvial cones has occurred in different phases, separated by a phases of fluvial erosion. In addition, the uniform, small clast sizes of the colluvial deposits and the absence of larger clasts implies a brittle condition of the parent rock, possibly by tectonic stress during the various stages of orogeny. Finally, the observed lithologies of the colluvial deposits do not match the geological map of the area and may reveal outcrops of Ordovician shales or Precambrian schists in the lower part of the slope.

A remarkable feature in the upper drainage basin close to the Abra de Lipán at an altitude of approximately 3,900 m.a.s.l. has been discovered during analysis of aerial-photography (Fig. 111). It shows up as a scar-like semi-circle of nearly 150 m in diameter opening downslope. The slope rock consists of Precambrian schists and phyllites, but it is areally covered with slope debris. Its aspect is roughly north and the slope angle has been estimated to approximately 35°. The vertical displacement of the scar does not seem to exceed very few meters. Particularly judging from its semi-circular shape and slight vertical displacement, this feature has to be interpreted as an initial landslide. Whether it incorporates massive rock or only applies to the uppermost layer of debris cannot be decided based on the available data. Nevertheless, this possible location of a future landslide demands further attention and should be of particular political interest as it is directly crossed by National Road No. 52.

With the exception of this example, no present processes of mass wasting or associated deposits have been observed in the study area. The geomorphological conditions of today do not seem to account for the above stated examples. Therefore, the process of landsliding and mass wasting has been apparently favoured by conditions different from today.
Based on the observations in the study area two explications seem reasonable and may be considered as possible scenarios for the formation of landslides and rock slides. Wetter

climatic conditions would certainly have had significant effects on slope stability. This is particularly true for the Ordovician slates, in which the two major landslides of study area have occurred. On the other hand, the rockslide observed in the Quebrada de Sepulturas does not require an explanation involving a climate change. Due to its geomorphological setting it may very well have been the sole result of lateral undercutting because of a drainage diversion by increased alluvial fan activity. In addition, the lithological and geologic slope characteristics point to a pre-weakened condition of the parent rock. Therefore even a single earthquake may have been a possible trigger for the rockslide.

In many cases, earthquakes and neotectonic activity in association with major thrust systems and active faults have been reported to have triggered rock slides and rock avalanches in NW-Argentina (GONZÁLES DÍAZ AND MON 1998, HERMANNS AND STRECKER 1999, STRECKER AND MARRETT 1999). A somewhat different explanation has been offered by TRAUTH AND STRECKER (1999) and TRAUTH ET AL. (2000) who correlate the avalanching events with phases of increased incision and undercutting resulting from significantly wetter climatic conditions. Evidently, both models could account for the observed mass wasting phenomena in the study area.

4.5. ALLUVIAL FANS

Alluvial fans are classified in a number of ways and several different dominant processes have been reported in the literature to be responsible for their formation (LECCE 1990, BLAIR AND MCPHERSON 1994). However, there is consensus that alluvial fans play an important role in dryland fluvial systems (BULL 1977, HARVEY 1997) and are therefore an essential issue for the geomorphic development of both the study area as well as all of NW-Argentina (CZAIJKA 1958B AND 1972, SCHÄFER AND SCHWAB 1975).

Within the study area mainly three different of alluvial fans have been identified based on their morphological characteristics and topographic correlation (table 5). Regarding the time involved in their development and the interpretation of paleoenvironmental conditions of their formation, the three types of alluvial fans have been assigned to different generations of fan growth with A-1 being the oldest phase and A-3 the youngest to recent phase.

GENERATION	LOCATION; MORPHOLOGICAL CHARACTERISTICS	Ø EXTENT [m]	ELEVATION [m.a.s.l.]	ACTIVITY	SLOPE [°]	FIG.
A-1	Lipán/Terraza Grande; deeply incised	600	3,100	No	8 - 11	112
A-1	Terraza Grande; no dissection or activity	250	3,000	No - rare	12	117
A-1	Qbd. de Sepulturas; isolated fan remnant	250	3,440	No	10	-
A-1	Qbd. de Estancia Grande; isolated fan remnant	200	3,400	No	10	116
A-1	Qbd. del Cobre; dissected fan remnant	150	2,850	No	18	113
A-1	Qbd. de Sunchoguaico; dissected, intercalated fans	1,000	2,750	Rare	6	114
A-1	Qbd. de Sunchoguaico; dissected, intercalated fans	300	2,700	No	19	115
A-1	Qbd. de Tumbaya; isolated, dissected fan remnant	100	2,580	No	10	-
A-1 / A-2	Qbd. de Lipán; no dissection, debris-flow deposition	200	3,250	Yes	11	101
A-2	Qbd. de Sepulturas; distal erosion by stream channel	900	3,300	No - rare	8	109
A-2	Qbd. de Estancia Grande; active debris-flow levees	500	3,100	Yes	8	118
A-2	Qbd. del Cobre; active debris-flows, distal erosion	700	2,620	Yes	?	119
A-2 / A-3	Qbd. de Est. Grande; debris-flow levee, multiphase	200	2,750	Yes	12	-
A-3	Qbd. de Lipán; incised into terrace slopes, multiphase	120	3,150	Yes	15	124
A-3	Qbd. de Lipán; single debris/grain flow event	70	3,100	Yes	20	123

Table 5: Examples of mapped alluvial fans and their characteristics.

- *Alluvial fans of generation A-1* are the largest fans within the study area. They have been observed in several quebradas at elevations between 2,700 and approximately 3,500 m.a.s.l.. In all cases, these fans are now elevated 40 to 200 meters above the present valley floor, while most of them have been deposited directly onto the terrace surface of terrace generation T-3. Particularly in the higher parts of the study area, where terrace deposits are absent, the alluvial fan remnants have been observed in no apparent context to any terrace. Only where located on lateral parts of terrace surfaces, they have been protected from erosion. Otherwise most fans have undergone severe dissection and destruction. Thus their morphological characteristics and size can sometimes only be reconstructed from the preserved fan remnants. These remnants of A-1 fans have been observed to have different sizes from a few hundred meters to more than one kilometer in length and radial extent.

Where an inactive but significantly inclined alluvial surface has been detected, but typical alluvial fan morphology was absent due to the advanced state of fluvial erosion and dissection, slope angle has formed the base for the recognition and classification of these remnants as eroded and dissected alluvial fans. The slope angles of all mapped A-1 alluvial fans in study area vary between 6° and 20°, while the fan remnants of larger extent show less inclined slope angles (table 5). In fact, contrary to river floodplains, dryland alluvial fans always show a relatively high inclination angle of up to 25° or more (BLAIR AND MC PHERSON 1994). Usually, two simple relationships have been noted to be characteristic of alluvial fan morphology: the larger the drainage area, the larger is the fan area and the smaller is the slope of the fan. Even though the advanced erosional state of some alluvial fan remnants in the study area did not allow the reconstruction of original fan sizes, the estimated extent and slope angles seem to confirm this inverse relation of alluvial fan slope and fan catchment area.

While most of the A-1 fans have only have preserved as remnants, the type of erosion responsible for the disintegration of the fans differs significantly. In the upper study area (e.g. Qbd. de Estancia Grande and Sepulturas) fan remnants stand about 30 to 60 meters above the present valley floor. They are very isolated, and direct topographic correlation to other fan remnants is difficult. However, the fan remnants have a relatively smooth morphology, they are entirely covered with grasses and shrubs and are not being subject to dissection by gullies or badland formation (Fig. 116). By contrast, fan remnants below approximately 3,000 m.a.s.l. do not only stand 100 to 200 meters above the valley floor, they are also subject to intense fluvial dissection and badland formation (Fig. 112). Sparse vegetation has been found only on the surfaces of these remnants.

These observations lead to several speculations. If present microclimatic differences between the upper and the lower study area are responsible for the different morphological appearances of the fan remnants, then their erosion must have been the result of paleoenvironmental conditions very different from the present conditions. Alternatively, if the present geomorphic processes responsible for the erosion and dissection of the fans remnants in the lower study area (badlands, gullies) were also responsible for the erosion of all fans in the past, paleoenvironmental conditions similar to the present conditions in the lower study area must have prevailed up to much higher elevations. This is not very likely as fan remnants do not show any sign of relict badland formation or gullying.

Finally, if erosion is supposed to work its way up the study area in a retrocedent way by headward extension, the first phase of erosion must have been interrupted for a significant length of time, subduing erosion and dissection and only recently allowing it to work its way up the study area again. However, this last scenario is not very likely, as the largest fan remnants have not been observed in the highest parts of the study area. Instead the remnants in the higher reaches are remarkably small and isolated, and are indicative of past erosional processes as intense as in the lower study area.

Fig. 112: *Deeply incised inactive alluvial fan of fan generation A-1 south of Lipán.*

Fig. 113: Deeply incised A-1 fan remnant in the Qbd. del Cobre.

Fig. 114: Interspersed fans of generation A-1 in the Qbd. de Sunchoguaico. Debris-flows and other fan shaping processes are still (again?) active on fan surface.

Fig. 115: Dissected fan segments of smaller fans (A-1) in the Quebrada de Sunchoguaico. Note how coloration indicates different catchment lithology and how all fans have adjusted to the same level.

Fig. 116: Entirely inactive fan remnants of generation A-1 in the upper Quebrada de Estancia Grande. Note the smooth appearance and absent dissection.

Fig. 117: Alluvial fan of generation A-1 on terrace surface at Terraza Grande showing no signs of dissection or activity.

Concluding from these considerations, the processes responsible for the intense erosion of the A-1 fans as well as the depositional terraces have likely been active under a geomorphic regime much different from today, most probably dominated by increased importance of fluvial processes and floodplain incision. Because many of the A-1 fans have deposited onto the terrace surface of terrace generation T-3, the erosion and dissection of the terraces have at the same time led to the exposure of the alluvial fan deposits. Therefore, these deposits have been included in the description and interpretation of the sedimentological profiles (4.3.2.1.) which has clearly recorded intense debris-flow action as the major process of A-3 alluvial fan deposition. However, in comparison to the underlying terrace deposits a significant decrease of average clast size, maximum clast size and the thickness of debris-flow layers points to an altered geomorphic regime, possibly in relation with paleoenvironmental change.

- *A-2 generation* fans are large alluvial fans with an extent of several hundreds of meters to over a kilometer. Again, their slope angle varies with extent and ranges from 8° to 15°. A-2 fans have been observed in the entire study area up to elevations of 3,800 to 3,900 m.a.s.l.. They are particularly common in the Quebrada de Estancia Grande, where the individual fans interfinger and cover almost the entire width of the valley (Fig. 118). Contrary to the A-1 fans, these fans have not experienced the same amount of intense erosion. Instead, in most cases they are practically adjusted to the present valley floor. Locally, lateral erosion on the floodplain has eroded the distal parts of the fan and created a scarp. These scarps are not very pronounced, mostly ranging from a few meters to 20 meters. For most of these fans, debris-flow levees and lobes on their surface are characteristic, otherwise the largest part of the fan surface is covered with vegetation like shrubs and cacti. These levees and lobes are characteristic for alluvial fans in drylands and result from concentration of boulders and larger clasts at the top, front and sides of debris-flows (COSTA 1984, BLAIR AND MCPHERSON 1998). Therefore these features point to an ongoing depositional activity of A-2 fans even though recurrence intervals are long enough to allow plant growth on inactive fan parts.

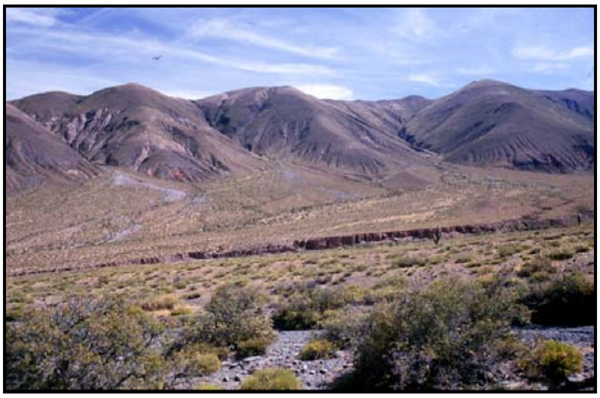


Fig. 118: Medium scale interspersed fans in the Quebrada de Estancia Grande. Note active alluvial fan lobes (left) and lateral erosion of drainage channel in distal fan areas (center).

Fig. 119: Medium-sized fan (Quebrada del Cobre). Note active debris-flow deposition (light grey area).

From their topographic as well as morphological characteristics A-2 fans are inferred to be much younger than A-1 fans. Particularly, their adjustment to the valley floor, their rather undissected morphological form and shape as well as their ongoing activity account for their younger age. In fact, the two fan generations must have been separated by severe erosion and dissection which has affected A-1 fans as well as great parts of the mountain slopes, but has done no harm to the A-2 fans (Fig. 120). Judging from their similar size and extent and supposing similar rates of debris-flow deposition, the duration of A-2 fans may approximately correspond to the duration of A-1 deposition.

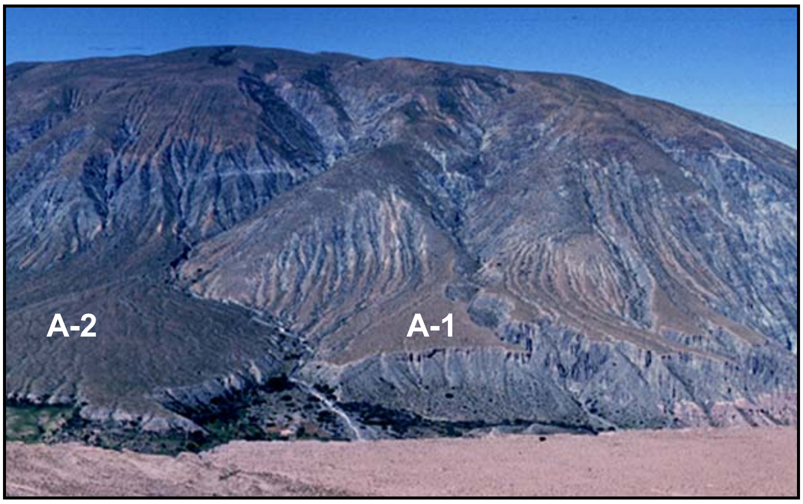


Fig. 120: Association of A-1 and A-2 generation alluvial fans in the Quebrada de Estancia Grande. Note the advanced state of erosion and dissection of the A-1 fan indicated by multiple drainage channels cutting it. In contrast, the A-2 fan is not dissected or cut by these drainage channels, instead it seems to cover them.

- *A-3 generation* fans are relatively small alluvial fans contributing sediment to the floodplains of the Quebradas from equally small drainage basins (Fig. 122). All A-3 fans are between a few tens of meters and up to more than one hundred meters long, their radial extent does not exceed a few tens of meters and their slope angle varies between 10° and 25°. In all cases the fan deposits consist of matrix poor, clast-supported fanglomerates pointing to deposition by debris-flows (Fig. 121). The average clast size ranges between three to five centimeters. Maximum clast sizes do not exceed 15 to 20 centimeters. The matrix appears to be a mixture of sand, silt and clay.

Fig. 121: Sedimentological characteristics of A-3 fan. Note fanglomeratic composition of deposit and marked layering (numbers) pointing to debris-flow deposition.

Fig. 122: A-3 fan in the Quebrada de Lipán; note ho the fan has buried trees while being deposited onto the floodplain.

Particularly on the fan surfaces, very little amounts of fine grain sizes are exposed, most probably a direct result of secondary winnowing by overland flow processes. The lithological composition of the fan deposits varies, depending on the characteristics of the fan catchments, but most fans have been observed to originate from Precambrian schists and phyllites as well as from fanglomeratic terrace deposits in badland areas. However, the fanglomeratic sediments show a clear layering within most of these A-3 fans. The layers of 40 to 60 centimeters thickness are separated by a concentration of coarse clasts and very little amounts of matrix, similar to the present surface. Considering the sedimentological characteristics of the sediments, these layers seem to indicate a polyphase formation of these fans by several debris-flow events having followed each other. Compared with the above described terrace deposits, these debris-flow deposits show similar textural characteristics, but clast size as well as thickness of each layer are up to an order of magnitude smaller than within most terrace deposits. This implies similar mechanisms of debris-flow deposition, but very different intensities of these processes, most probably owing to different environmental conditions.

Fig. 123: Small lateral fan in the Qbd. de Lipán with catchment of mainly schists of Precambrian Puncoviscana Formation. Note lobe-like fan morphology due to little content of matrix and secondary winnowing of fan surface ("sieving").

Fig. 124: Small lateral fan in the Quebrada de Lipán with catchment area of mainly Quaternary fanglomerates. Note fan trenching into relict colluvial slopes.

A small number of observed A-3 fans have shown a somewhat different morphology. While having an extent similar to most other A-3 fans, their surface is divided into irregularly spaced, few meter wide lobes (Fig. 123). These fans consist of matrix-poor fanglomerates, clast sizes average five to ten centimeters and maximum clast size does not exceed 15 centimeters. Corresponding to their catchment in highly dissected slopes of the Precambrian Puncoviscana Formation, their clast lithology is exclusively composed of schists. In any case, their formation may be attributed to processes similar to grain flows due to their irregular surface and the little amount of matrix (LOWE 1979). Alternatively, a process described as "sieving" may have removed most matrix during transport which has led to the fast deposition of the flow (HARVEY 1997).

A-3 fans occur in the entire study area and have been observed up to elevations of at least 3,400 m.a.s.l.. Even though in most cases A-3 fans are located in lateral parts of the floodplain, they have been observed in very different associations. While some fans have incised into colluvial slopes and badland areas, others have been found to bury trees and fences on the floodplain (Fig. 122). Some fans have been subject to sparse plant growth and lateral erosion by floodplain activity, others appear to be freshly deposited. Due to their small extent they should be of relatively young age. As they have affected older and larger landforms by burial or dissection and show evidence for present activity, T-3 fans are interpreted as corresponding to the youngest, recent to present phase of deposition within the study area.

Summing up the observed characteristics of the alluvial fans in the study area (Table 5), three generations of alluvial fans have been recognised. Considered an integrative part of the semi-arid morphodynamic system, the development of all fan generations has been predominantly controlled by debris-flow deposition. Despite the similar processes responsible for their formation, the fan morphology differs significantly. In most cases, the topographically highest fans have preserved only in remnants, showing signs of enormous erosion and dissection. Therefore these fans (A-1) are interpreted as the oldest fan generation in the study area. The period of their deposition is supposed to have been followed by a period of intense fluvial erosion and dissection, affecting not only the alluvial fans but also the fluvial terraces onto which they have been deposited. While a few fans of the following generation (A-2) still adjusted to terrace remnants, most of these alluvial fans formed in adjustment to the present valley floor. Even though these fans are presently active, their size indicates that their activity extends relatively far back into the past. However, minor erosion has subsequently affected these fans, possibly corresponding to minor environmental fluctuations. The last and smallest generation of alluvial fans (A-3) is inferred to be of very young age. These fans are in most cases the result of ongoing slope dissection and badland formation, contributing remarkable amounts of material to the floodplain. Alluvial fan construction by debris-flow deposition has been an integrative part of the geomorphic system of this semi-arid region not only during long periods in the past, but it continues to be important for the present development study area.

4.6. PEDOLOGICAL INFORMATION

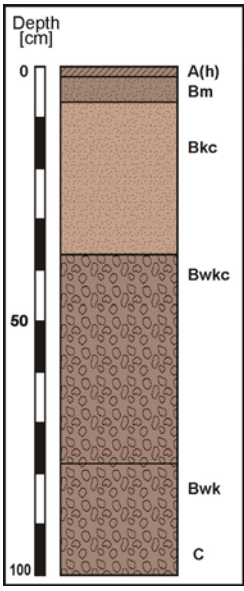
As soils develop in balance with the geoecological conditions of their environment, soil formation depends to a large extent on climatic and hydrological factors (SCHEFFER AND SCHACHTSCHABEL 2002). Particularly in high mountain environments, where geoecological conditions are subject to a strong hypsometric gradient, pedological information can add valuable data about paleoenvironmental conditions and their change over time. On the other hand, intense processes of erosion and sedimentation as well as the complexity of climatic factors usually cause soils in mountain environments to be relatively young having a high lithic component (EITEL 1999). Generally, this also seems to be true for the study area where pedogenic processes are slow. Only very few well-developed soil types have been reported from semi-arid NW-Argentina (2.6.). Furthermore, in most cases these soils seem to be the product of past environmental conditions and must have formed on landscapes of the past, which makes them paleosols by definition (MACK ET AL. 1993).
In this chapter, all pedological information and observations from the study area will be summarized and briefly discussed. Corresponding to the intense present geomorphic activity in large parts of the study area, soils and paleosols have exclusively been observed on top of the fluvial terraces (T-1 to T-3) and on the surfaces of the oldest alluvial fan generation A-1. While some soils and their genetic processes are still evident from their typical association of horizons in a vertical profile, others have to be inferred from micromorphological analysis because only indivdual soil components are left (for localities of soil samples see figure 205 in the appendix). Therefore the pedological profiles and the micromorphological analysis of significant soil components will be discussed seperately.

4.6.1. SOIL PROFILES AND SOIL TYPES

The description of the following soil profiles (Fig. 125-136) is based on two different types of profiles. Locally, erosional processes have cut into sediments and soils. This allows pedological analysis up to several meters of depth. Otherwise, the high amount of clasts in fanglomeratic parent material has complicated pedological field analysis enormously. Due to these difficulties this chapter aims to draw very general conclusions from the pedological observations instead of giving a complete description of the soils of the study area.

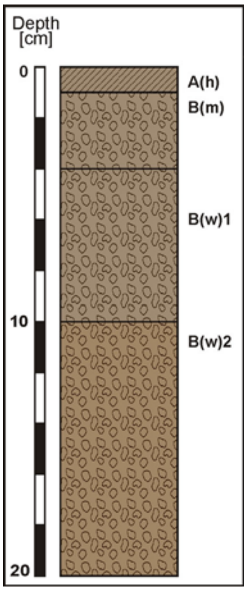
SLI-1 The soil profile SLI-1 was taken on top of depositional terrace T-3 in close vicinity to a drainage channel. The profile is characterized by a thick unit of a cemented sandy fanglomerate overlying fanglomeratic material with markedly reddish matrix. Overall $CaCO_3$-content is relatively high and decreases downwards. This should roughly imply the following sequential development: soil formation (reddish coloration) - sand accumulation - subsequent cementation - carbonatization.

SLI-2 The soil profile SLI-2 was taken on the subhorizontal flat top of depositional terrace T-3 without any drainage channels nearby. The profile encompasses only the top 20 cm beneath the surface and mainly consists of fanglomeratic material. It does not show any sign of pedogenesis except a slightly reddish color of the matrix and a very low $CaCO_3$-content.



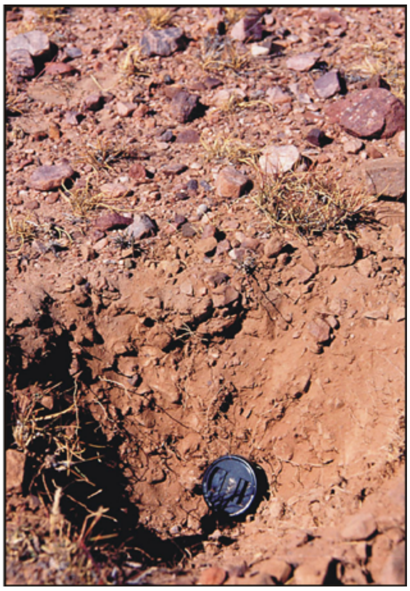
SLI-1
65,578W 23,658S

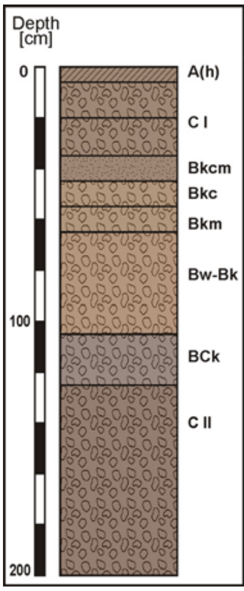
Horizon	Description
A(h)	sparse vegetational cover (shrub, grass bunches), pebble pavement
Bm	sandy, fine fanglomerate (rounded to subrounded clasts) yellowish (5YR 5/4) no $CaCO_3$-reaction, cemented
Bkc	sandy, fine fanglomerate to conglomerate (5YR 7/4), uncemented moderate to strong $CaCO_3$-reaction some mm-carbonate nodules and coatings
Bwkc	fine fanglomerate, reddish color (10 R 5/4), sandy to silty matrix moderate $CaCO_3$-reaction, few carbonate nodules and coatings
Bwk	fine fanglomerate, reddish color (10 R 5/4), sandy to silty matrix weak to no $CaCO_3$-reaction, very few carbonate coatings
C	



SLI-2
65,578W 23,657S

Horizon	Description
A(h)	very little to no vegetational cover (grass), pebbles
B(m)	sandy, fine fanglomerate, slightly reddish (10YR 5/4) weak to no $CaCO_3$-reaction, some roots, slightly cemented
B(w)1	sandy, fine fanglomerate, slightly reddish (10YR 5/4) weak to no $CaCO_3$-reaction, some roots
B(w)2	sandy, fine fanglomerate, reddish (10YR 5/6) weak to no $CaCO_3$-reaction, some roots



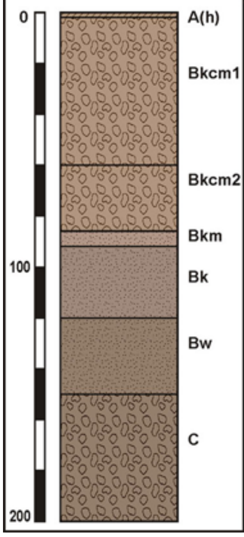


SLI-3
65,584W 23,657S

Horizon	Description
A(h)	sparse vegetational cover (shrub, grass bunches), little organic material
	clast-supported fanglomerate with silty-sandy matrix (7,5YR 5/4) no $CaCO_3$-reaction, roots
C I	matrix-supported finer fanglomerate with silty-sandy matrix (7,5YR 5/4) no $CaCO_3$-reaction, some roots
Bkcm	sandy, very fine fanglomerate, cemented (7,5YR 5/4) $CaCO_3$-reaction on roots, nodules and clast coatings
Bkc	fanglomerate with matrix (10YR 6/4) strong $CaCO_3$-reaction, some carbonate precipitates
Bkm	fine fanglomerate, cemented matrix (10YR 6/4) $CaCO_3$-reaction with matrix
Bw-Bk	coarse clast-supported fanglomerate, reddish matrix (5YR 6/4) $CaCO_3$-reaction with matrix
BCk	fine clast-supported fanglomerate, little matrix (2,5YR 5/2) weak $CaCO_3$-reaction with matrix
C II	pebbly fanglomerate, very little matrix (2,5YR 4/4) no $CaCO_3$-reaction

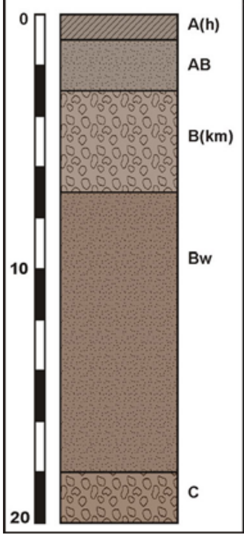
Fig. 125-127: Soils of the study area (SLI-1, SLI-2, SLI-3).

STG-1
65,568W 23,673S



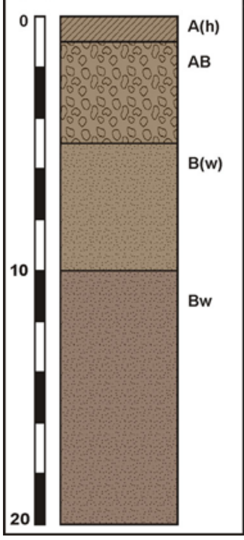
Horizon	Description
A(h)	sparse vegetational cover (shrub, grass bunches), pebble pavement
Bkcm1	coarse fanglomerate (7,5YR 6/4) strong $CaCO_3$-reaction with matrix, pebble coatings, cemented
Bkcm2	clast-supported fanglomerate (7,5YR 8/4) moderate $CaCO_3$-reaction with matrix, pebble coatings, cemented
Bkm	sandy fine fanglomerate (5YR 5/3) weak $CaCO_3$-reaction on coatings and roots, cemented
Bk	sandy fine fanglomerate (5YR 6/3), many roots weak $CaCO_3$-reaction on roots
Bw	sandy fanglomerate, some floating pebbles, reddish (7,5YR 4/4) no $CaCO_3$-reaction
C	matrix supported fanglomerate (7,5YR 4/4) no $CaCO_3$-reaction

STG-2
65,561W 23,682S



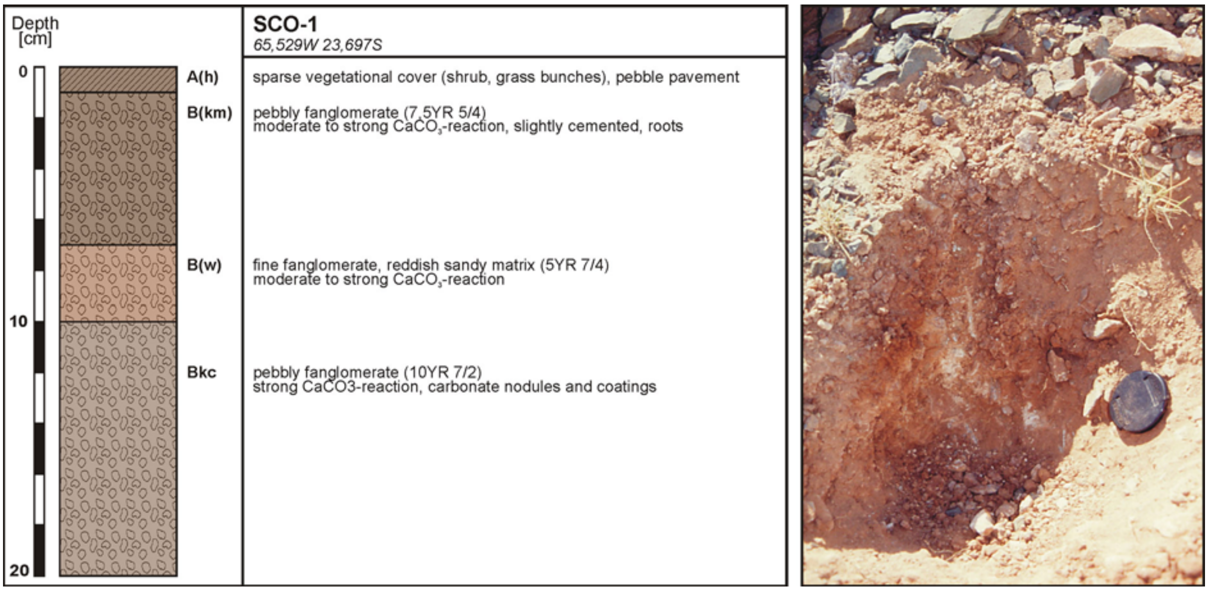
Horizon	Description
A(h)	sparse vegetational cover (shrub, grass bunches), pebble pavement
AB	unconsolidated sand, some pebbles (10YR 5/2) no $CaCO_3$-reaction
B(km)	fine fanglomerate, sandy to silty matrix (7,5YR 6/2) weak $CaCO_3$-reaction on pebble coatings, slightly cemented, some roots
Bw	silty-sandy, clay rich material, some pebbles, reddish color (5YR 4/4) no $CaCO_3$-reaction, clay forms aggregates
C	fine fanglomerate, sandy to silty matrix (7,5YR 5/4) no $CaCO_3$-reaction

SPG-1
65,589W 23,673S



Horizon	Description
A(h)	sparse vegetational cover (shrub, grass bunches), pebble pavement
AB	sandy, fine fanglomerate (10YR 5/4) weak to no $CaCO_3$-reaction
B(w)	sandy to silty material, many pebbles (10YR 5/4) weak to no $CaCO_3$-reaction
Bw	silty and clay-rich matrix (5YR 4/4), some pebbles no $CaCO_3$-reaction, roots, clay aggregates

Fig. 128-130: Soils of the study area (STG-1, STG-2, SPG-1).



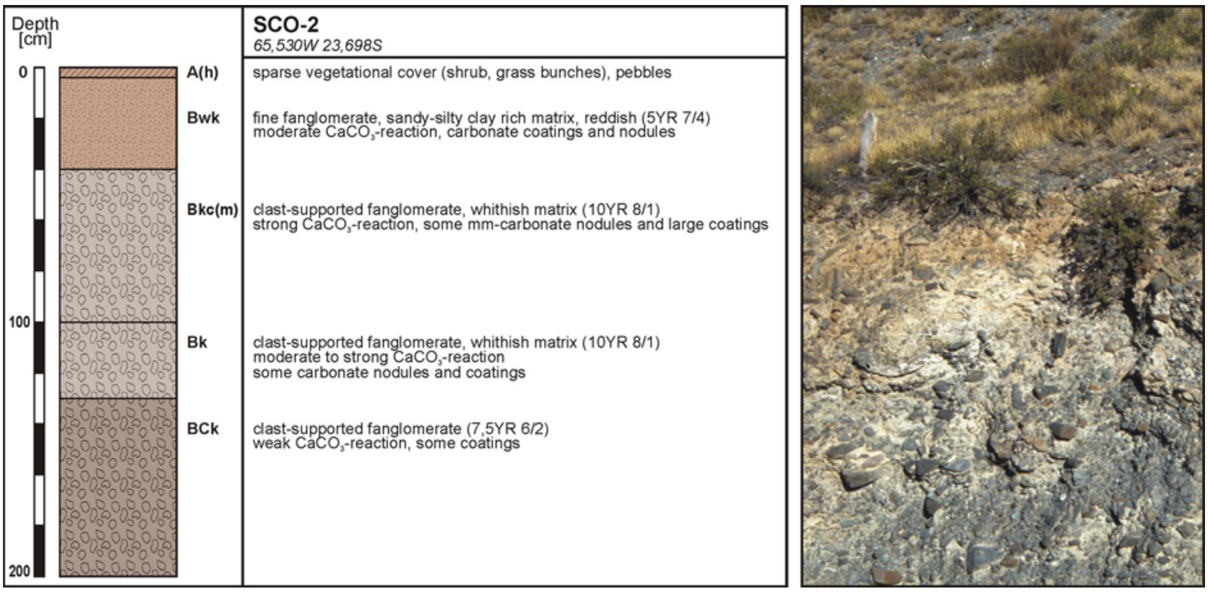


Fig. 131-133: Soils of the study area (SPG-2, SCO-1, SCO-2).

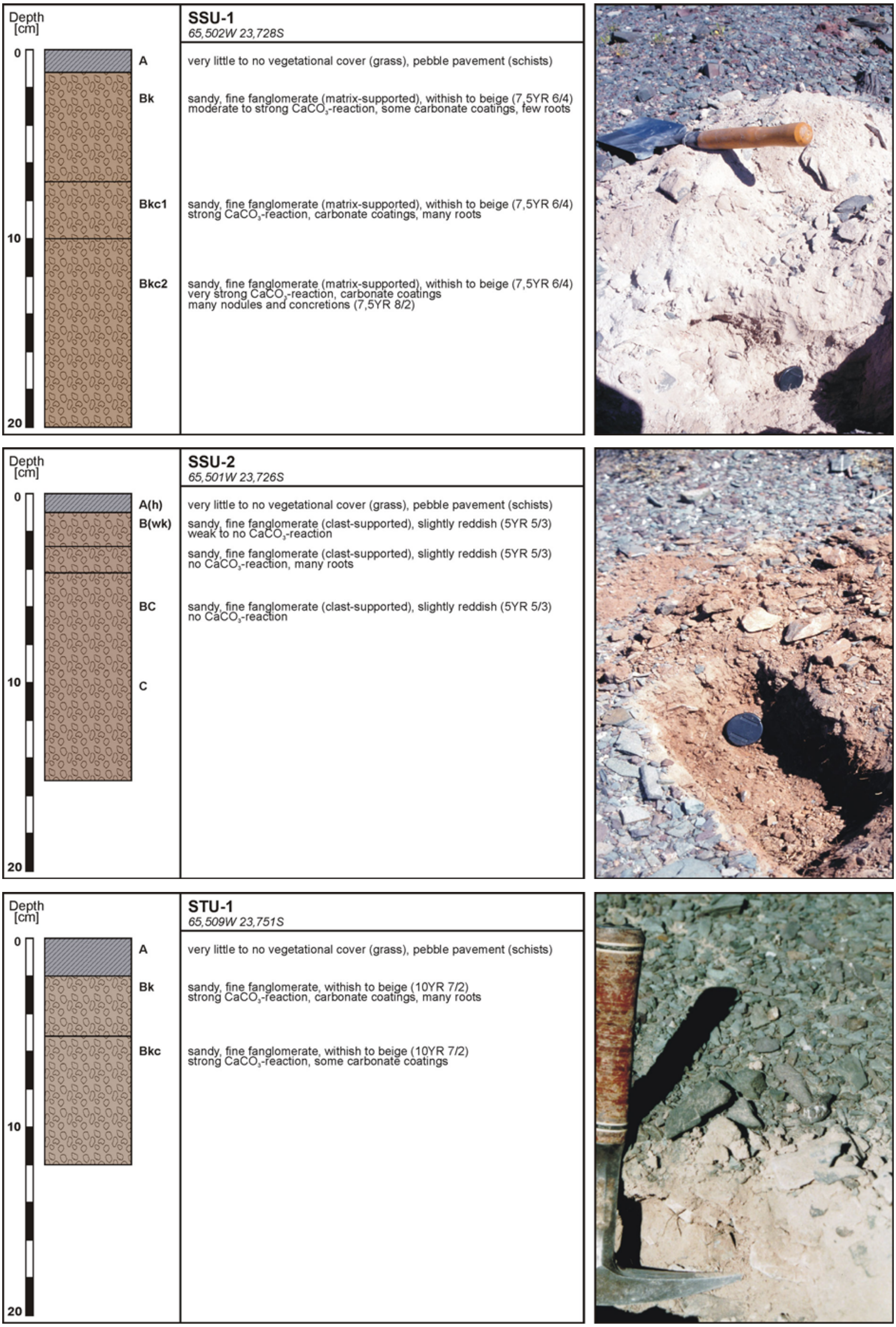


Fig. 134-136: Soils of the study area (SU-1, SU-2, STU-1).

This should imply very weak to no soil formation. However, all neighbouring deeper profiles do show signs of soil formation, suggesting that initial soils are characteristic only for the topmost horizons and might by underlain by older well-developed horizons (e.g. SLI-1, SLI-3, STG-1).

SLI-3 The soil profile SLI-3 was taken on top of the depositional terrace T-3 in an active drainage channel. Its topmost part consists of fanglomeratic material without any signs of soil formation which overlies two horizons of cemented sandy material. Below, a reddish horizon grades into the fanglomeratic parent material. While the topmost part is essentially free of carbonate, the B-horizon has a high $CaCO_3$-content, which decreases downwards. For the sequential development of the soil the following picture can be drawn: soil formation (reddish coloration) - sand accumulation - subsequent cementation – carbonatization – recent to present accumulation.

STG-1 The soil profile STG-1 was taken on top of an A-1 alluvial fan within a small drainage channel. Its topmost part horizon is characterized by a well-cemented fanglomerate overlying a sandy cemented horzion. Below, the sandy horizon grades into a reddish fanglomerate. The profile has a high $CaCO_3$-content in the upper horizon decreasing downwards. Concluding from the above this should imply the following sequence of soil development: soil formation (reddish coloration) - sand accumulation – subsequent weak cementation – accumulation – carbonatization and cementation.

STG-2 The soil profile STG-2 was taken on top of the flat, subhorizontal surface of the depositional terrace T-3 at Terraza Grande. Even though it encompasses only the top 20 cm beneath the surface it shows a well-developed B-horizon characterized by markedly finer grain sizes and a striking reddish coloration. The entire profile is free of significant amounts of $CaCO_3$. However, it gives clear evidence for a phase of strong soil formation. The timing of this phase can not be inferred from the profile although the scarcity of vegetation as well as the poorly developed A-horizon point to a relict character of the B-horizon.

SPG-1 The soil profile SPG-1 was taken on top of the flat, subhorizonal surface of the depositional terrace T-1 at Potrero Grande. It is dominated by a well-developed B-horizon of fine grain sizes and reddish coloration. The entire profile is essentially carbonate free. Similar to profile STG-2 a phase of strong soil formation, most probably under very different conditions than the present ones, seems to be responsible for the intense development of the B-horizon.

SPG-2 The soil profile SPG-2 was taken on top of the flat, subhorizontal surface of the depositional terrace T-2 at Potrero Grande. It is dominated by coarser grain sizes and a uniform color. There are abundant fragmentated carbonate clasts in the lower part. The overall $CaCO_3$-content of the profile is very high. Regarding these characteristics a very strong phase of carbonatization and cementation must have preceded fragmentation. Therefore the profile does not seem to have formed under present conditions.

SCO-1 The soil profile SCO-1 was taken on top of an A-1 alluvial fan in the Quebrada del Cobre. It is developed in a pebbly fanglomerate interrupted in its middle part by a reddish horizon. The $CaCO_3$-content is highest in the lowermost horizon which shows strong signs of carbonatization. Whether strong carbonatization predates or postdates the formation of the reddish horizon or is even linked to it, remains unclear.

SCO-2 The soil profile SCO-2 was taken on top of an A-1 alluvial fan in the Quebrada del Cobre within a drainage channel. It is dominated by a thick lowermost horizon of high $CaCO_3$-content underlying a thin reddish horizon of moderate $CaCO_3$-content. Similar to profile SCO-1 the sequential development of this profile remains unclear without understanding the context of soil formation leading to the thin reddish horizon. However, the fact that the carbonate horizon weakens downwards and carbonatization incorporates the reddish horizon points to a younger age of the carbonatization phase relative to the reddish horizon.

SSU-1 The soil profile SSU-1 was taken on top of an A-1 alluvial fan in the Quebrada de Suncho-guaico. It is characterized by a whitish powder-like horizon of sand to silt size material of extraordinary high $CaCO_3$-content (carbonate dust) underlying a desert pavement of mainly Precambrian schists. Whether there is a genetic link between the rock type of the desert pavement and the mode of accumulation of carbonate dust cannot be decided on the base of the available data. In any case the small amount of cementation and lack of evidence for other soil formation implies a relatively young age of these carbonate accumulation horizons.

SSU-2 The soil profile SSU-2 was was taken on top of an A-1 alluvial fan in the Quebrada de Sunchoguaico. It shows a slightly reddish and very weakly developed soil horizon. Similar to profile SLI-2 the small profile depth in combination with the weakly developed horizon points to present conditions not very favourable for soil development. Judging from the existent data, the possibility of an undetected deeper paleosol can not be excluded.

STU-1 The soil profile SSU-1 was taken on top of an A-1 alluvial fan in the Quebrada de Tumbaya. It is characterized by a whitish powder-like horizon of sand to silt size material of extraordinary high $CaCO_3$-content (carbonate dust) underlying a desert pavement of mainly Precambrian schists. Similar to profile SSU-1 the exact conditions and the relative timing of the accumulation of carbonate dust remains unclear.

From the soil profiles described above, particularly four contrasting types of diagnostic horizons are remarkable. These are horizons of weak to essentially no sign of soil formation, reddish and clay-rich horizons mostly in the lower part of the profiles, carbonate-rich to carbonate-cemented horizons, and very sand-rich and cemented horizons. All of these horizons point to pedogenic processes very different from each other, assigning a relict and polygenetic character to most observed soils.

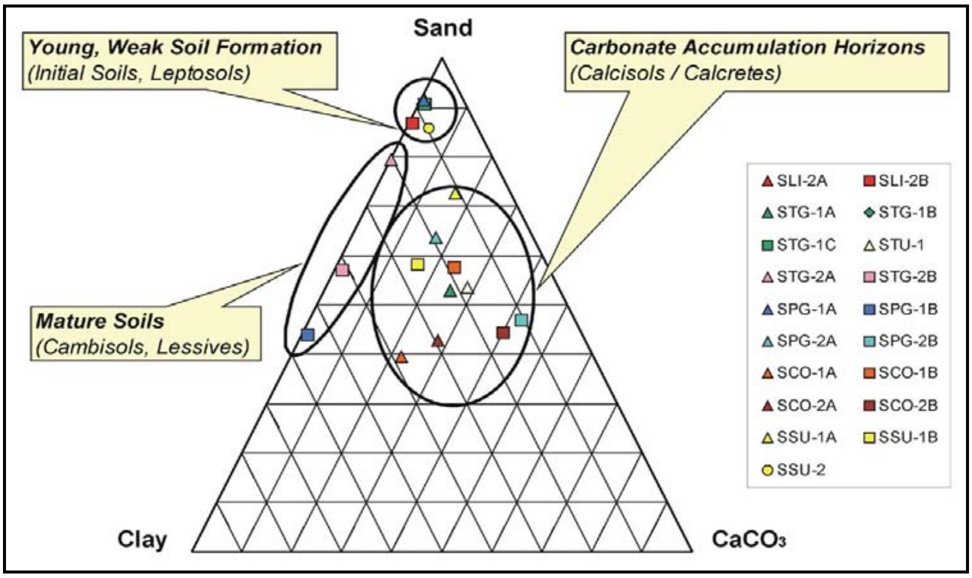


Fig. 137: Definition of different soil types in the study area on the base of grain size and $CaCO_3$ data. For data and location of the soil samples see appendix.

Nevertheless, each diagnostic horizon is indicative of a certain pedogenic process. Depending on how intense the characteristics of each diagnostic horizon have developed, the denomination of the different soils types varies from one case to the other. To differentiate between the various types of soil horizons and to compare them regarding the intensity of their development, 19 samples from ten profiles were taken. Because grain size and $CaCO_3$-content are among the most commonly used pedological parameters indicative of soil formation (SCHEFFER AND SCHACHTSCHABEL 2002), resulting data for grain size composition and $CaCO_3$-content offer a good possibility to separate the different horizons and their associated pedogenic processes within the study area from each other. From figure 137 three of the four diagnostic horizons from field observation can be identified and differentiated. Their characteristics lead to the following definition of soil types and associated pedogenic processes (denomination after FAO 1990):

Leptosols

This group of soils could be described as initial soils of very weak to no soil formation. The soils are characterized by a dominance of coarse, pebbly to sandy grain sizes which is possibly the result of removement of the fines by percolationg rain water or interflow. $CaCO_3$-content is usually low. In most cases soil formation seems to be hindered by the lack of vegetation and/or the existence of a pebble pavement preventing water from infiltration. At many places these soils have been observed only superficially overlying older soils and soil horizons. This implies that these soils are relatively young, possibly even formed under present environmental conditions. Soils with the characteristics described above could be defined as *leptosols* (e.g. soil profiles SLI-2 and SSU-2 from the study area).

Cambisols and Luvisols

This group of soils is characterized by an intensely reddish colored horizon with a significant increase of clay and silt. A similar soil type has been reported from several locations in NW-Argentina (WERNER 1971, SEGEMAR-ITGE 1998, KRISL 1999). Reddish color seems to indicate rubefication resulting from the oxidation of hematite (Fe_2O_3), a process typically known from areas of intense summer drought (EITEL 1999). Rubefication could also account for the high content of clay and silt. A reddish soil with dominant rubefication could be defined as *chromic cambis*ol (e.g. soil profiles STG-2 and SPG-1 constitute from the study area).

Increased clay content could also be the result of translocation processes washing the clay out of the upper horizon into lower ones. This is the case in *luvisols,* which have been reported from some areas in NW-Argentina (ZIPPRICH ET AL. 1999), sometimes in association with a carbonate accumulation horizon resulting from the translocation of carbonate from the upper horizon (WERNER 1971, EBERLE 2000). However, it seems likely that climatic conditions required for the translocation and mobilization of clay would have made an enrichment of carbonate in the lower soil horizons impossible. Therefore the pedological association of a clay-rich horizon on top of a carbonate-rich horizon implies a polygenetic character of the soil. Carbonatization has been followed by clay translocation, both induced by different environmental conditions (BUSCHIAZZO 1985, MACHETTE 1985).
For both, the formation of cambisols and luvisols the soil has to be essentially carbonate free. Therefore, horizons with signs of carbonate accumulation on top of clay-rich horizons (*cambisols, luvisols*) as observed in the study area would imply a process of carbonatization much younger than the formation of cambisols and luvisols (e.g. soil profiles SLI1, SLI3 and STG1).

Calcisols

For this group of soils a high $CaCO_3$-content is characteristic. While grain size varies significantly, all soils with a horizon of more than 15 % $CaCO_3$-content within the uppermost 125 cm are called calcisols (FAO 1997). A petric calcisol shows additional signs of cementation within its carbonate-enriched horizons. The nongenetic term calcrete (S*panish: caliche*) stands for a cemented, near surface horizon of accumulated $CaCO_3$, it represents a distinct horizon *within* a soil profile and is not equal to a soil type (MACHETTE 1985, WRIGHT AND TUCKER 1991). Pedogenic calcretes refer to indurated and cemented calcic soils (WRIGHT AND TUCKER 1991). While cementation is very often a result of subsequent exhumation, dessication and erosion of the topmost soil horizons leading to the surficial preservation of these carbonate crusts (EITEL 1999), a variety of different models for hardening have been proposed, including biogenic activity (e.g. VERRECCHIA 1994). Carbonate crusts have been observed frequently and in different associations within the study area and will be discussed below (4.6.2.).
However, in many soils of the study area the process of carbonatization is evident from a variety of features. In most cases the $CaCO_3$-content ranges from 15 to 30 % while a few samples show values of up to 40 to 60 % of $CaCO_3$-content. In addition, carbonate nodules and concretions as well as pebble coatings indicate an ongoing carbonatization. The different geomorphological and microclimatic settings of each soil cause different types of carbonatization and therefore different appearance of these calcic horizons. An enormous number of literature proposes several models for the process of carbonatization and suggests several possible classification schemes for calcic soils and calcretes (WRIGHT AND TUCKER 1991). The simplest way uses morphological criteria of the calcic horizon to differentiate between various soil and calcrete types (WRIGHT AND TUCKER 1991) and leads to the following definition of calcic soils within the study area:

- *Calcified soils* are loosely to firmly cemented soils with carbonate accumlations greater than 10 %. They might show grain or pebble coatings and small nodules. In the study area soil profiles SLI-1, SLI-3 and STG-1 show these characteristics in their upper horizons while their lower horizons are increasingly free of carbonate and could be described as a cambisol. Thus, they apparently correspond to calcaric cambisols.

  In some soil profiles (SCO-1 and SCO-2) a thick horizon of carbonate accumulation (25 - 45 % $CaCO_3$–content) appears to underly a thin reddish B-horizon of moderate $CaCO_3$-content (15-20 %). The reddish horizon might correspond to a horizon of leaching, in which the carbonate gets trapped and from which it percolates laterally or downwards (WERNER 1971), possibly due to strong soil water movements on the steep slopes of SCO-1 and SCO-2. These soil profiles would generally correspond to calcisols. Only where the carbonate accumulation horizon is subject to subaerial exposition it appears hardened (WERNER 1971, BUSCHIAZZO 1985), e.g. at erosional channels, gullies or at the soil surface (Fig. 139).

- A fine, usually loose powder of mainly $CaCO_3$ as a continuos body with little nodule development is termed *powder calcrete*. Soil profiles STU-1 and SSU-1 correspond to this type of calcisol. EITEL (1999) attributes the powder-like material to $CaCO_3$-preciptiation but does not make any suggestion regarding the mechanisms of precipitation or the source of $CaCO_3$.

- A *hardpan calcrete* corresponds to a petrocalcic horizon within a soil profile. It usually has complex internal fabrics and structure, a sharp upper surface and a sheet-like appearance. The soil profile SPG-2 shows these characteristics with the exception that no sharp boundaries exists and calcrete occurs in fragments within a sandy carbonte matrix. This might imply a brecciation of an older generation of calcrete.

4.6.2. CARBONATE CRUSTS AND CALCRETES

As discussed above, calcic soils occur in a variety of different morphological forms and types, while calcretes are by definition cemented and hardened horizons. Nevertheless, they do form in different settings and phases due to a variety of processes which makes them in several ways polygenetic features (VERRECCHIA 1994).
WRIGHT AND TUCKER (1991) present a possible classification scheme based on the hydrologic setting of the calcrete. They differentiate pedogenic calcretes within a soil profile from non-pedogenic calcretes. Non-pedogenic calcretes can occur surficially, particularly at morphological steps and scarps like gully beds, where they are the result of sheetwash and interflow processes (*"Hangwasser-Carbonatisierung"* after SCHEFFER AND SCHACHTSCHABEL 2002). Within the study area non-pedogenic calcretes have been observed in a variety of different settings (fig. 138-141). While in many cases this classification does not reflect the genetic history of carbonatization of the calcrete, it might give hints to the mechanisms of cementation and hardening.

Fig. 138: Outcrop of calcrete on top of T-2 terrace surface at Potrero Grande.

Fig. 139: Detail of Fig. 138. Note roughness and fragmented appearance of calcrete.

Fig. 140: Outcrop of calcrete at terrace scarp at Terraza Grande. Note color differences due to varying source material within underlying fanglomerates.

Fig. 141: Detail of Fi.g 140. Note thickness of the multiple horizons showing carbonate accumulation.

Outcrops of calcretes have been observed mainly in two different settings within the study area. Calcretes have been found at the surface in close association to topographic irregularities like steps or ridges (Fig. 138). Presumably, the enhanced conditions for erosion by sheet flow have removed the upper soil horizons formerly having covered the calcrete. Alternatively, intensified runoff below a topographic step might redistribute dissolved carbonate and contribute to a continued carbonatization. Either of these assumptions is corroborated by the ovservations of calcretes at the surface of steep alluvial fan remnants (e.g. samples SCO-2, CC-11) where overland flow is expected relatively high due to the high inclination of 18°.

In addition, calcretes have been observed along the rims of dissected A-1 alluvial fans and terraces (Fig. 140 and 141). Here, an increased input of carbonate and subsequent hardening might be the result of carbonate saturated interflow. Thus, the interflow might recycle carbonate from older calcretes upslope. However, a closer look at these calcretes along the rims of the A-1 alluvial fans reveals several layers of calcrete on top of each other, separated by thin sheets of gravel. This points to alternating processes of carbonatization and sediment accumulation, possibly by overland flow or even sheet flooding. The calcrete layers have locally been observed up to five meters thick (Fig. 141) implying a relatively long and multiphase period of formation.

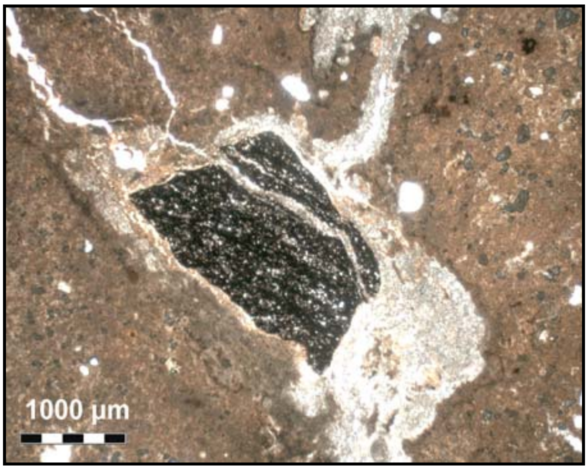
For further interpretation micromorphological analysis is essential and has formed the base for the definition of two end-members of microstructures in petrocalcic soils and calcretes reflecting the predominant type of cementation and calcite crystallization (WRIGHT 1990). The α-tpye resembles a calcrete mainly having formed due to evaporation-driven processes and the β-tpye originates from biogenic activity. Both types exhibit a variety of typical micromorphological features. Therefore, besides very general methods for micromorphological analysis of thin sections (e.g. ADAMS AND MCKENZIE 1998), the following description of five thin sections extracted from samples of calcrete within the study area refers to the applied scheme of WRIGHT (1990) and WRIGHT AND TUCKER (1991).

S-14 This sample was extracted from a carbonate fragment in the lowermost horizon of the soil profile SPG-2 on top of terrace T-2 (*65, 58W 23,674S*). The fragment seems to show signs of weathering, it appears dark greyish and its surface is occupied by roots.
The thin section shows few rock clasts and mineral fragments supported by a light ochre-brownish matrix. Rock clasts are mainly phyllitic and seem to be subject to cracking in some cases. The mineral components seem well-rounded, they are mainly quartz, kalifeldspar, plagioclase and some well-rounded hornblende and biotite. The very fine, dense, cryptocrystalline matrix is carbonatic. It is ochre to light-brownish but shows some variations of coloration. The matrix is interspersed with pores and fissures. Many of these have very rounded shape, possibly pointing to karstification. They are mostly filled by a whitish, fine carbonate matrix (crystallaria), which is not as fine-grained as the brownish matrix. Locally, the filling is not complete. Often, rock and mineral fragments seem to be spherically mantled by this type of whitish matrix. At many parts the crystallaria are surrounded by dark, spotted lines, possibly resulting from manganese precipitation. Very often slight traces of a very fine, orange material are found inside the crystallaria, possibly illuviated clay.

CC-14 This sample was collected from the surface of terrace T-2 just below a topographic break (possibly the result of an earthquake) of approximately four meters hight *(65,577W 23,677S)*. The sample appears in a whitish color a does not show any signs of weathering. Instead its surface exhibits a smooth, laminar aspect.
The thin section shows some rock clasts and minerals floating in a brownish-greyish matrix, which is frequently discrupted by pores and fissures. Rock clasts are of several origins, e.g. a hematite bearing quartz-muscovite-phyllite, some metapelite, metagreywackes and quartzites. The poorly sorted mineral components are mainly quartz and plagioclase, but also some epidotes, turmalines, hornblendes and to a lesser extent kalifeldspars and opaque fragments. The cryptocrystalline carbonatic matrix is light brownish to greyish, disrupted by irregular cracks and fissures. Even the finest cracks are subject to calcitic crystallization. Not everywhere the rock fissures are have been completely filled up. In addition, abundant nodules and peloid features can be observed. Mostly, these are round, spherical features with a nucleus of matrix fragments, surrounded by bended, thin structures. These thin structures are not always associated to peloids but can be observed as individual elongate features within the matrix.
In its lower part the thin section shows a sharp transition to a very different fabric of abundant mineral grains with orange matrix (compare samples CS-17 and CS-18 in 4.6.3.). Just above, cracks and fissures are extraordinary frequent and tend to be surrounded by dark, spotted lines, possibly manganese precipitates. Towards the upper part (surface) of the section these lines gradually decrease. Instead the upper part is characterized by a clear laminar fabric of little clasts and mineral fragments and contrasting coloration. Partly, very fine grained orange material is incorporated into the matrix.

CC-5 This sample was collected from the rim of Terraza Grande (terrace T-3), close to soil profile STG-1 *(65,568W 23,671S)*. It appears whitish to greyish, relatively unweathered and seems rather porose.
The thin section shows a dense matrix with abundant rock and mineral fragments. Among the rock fragments quartzite, muscovite-quartz-glimmer-schists and some metapelites have been observed, while mineral components are mainly quartz, some plagioclase and to lesser extent epidote and hematite. They are very angular to very little rounded. The cryptocrystalline matrix is brownish and shows several cracks. These cracks are partly filled with a lighter, greyish matrix, mantling rock and mineral fragments as well as fragments of the brownish matrix. In caverties and at the walls of the cracks orange colored fine-grained material has accumulated, possibly being illuviated clay. The entire section is rich in black pebbles, which seem to be organic material mantled by carbonatic matrix.

CC-11 This sample was collected from a surficial outcrop on top of an A-1 alluvial fan at the Quebrada del Cobre *(65,529W 23,698S)*. The fragmented sample appears whitish and porose with no weathering signs.
The thin section shows some rock clasts and mineral fragments. Rock fragments are quartzitic and metapelitic. Floating mineral fragments are not very abundant and exhibit a uniform grain size. They are dominated by quartz and plagioclase but also some hornblendes, turmalines and opaque fragments. The cryptocrystalline matrix is generally brownish to greyish. Around rock fragments, color and texture changes and shows more contrast. There are abundant spherical features, possibly nodules and peloids. In some parts these features are relatively large, contain predominantly matrix material and are surrounded by cracks while others contain various smaller fragments of matrix material. In the entire section very fine, orange material is incorporated, particularly within the small amount of crystallaria. The matrix is characterized by huge color differences and dark, spotted parts occupy large parts, possibly due to manganese precipitation.



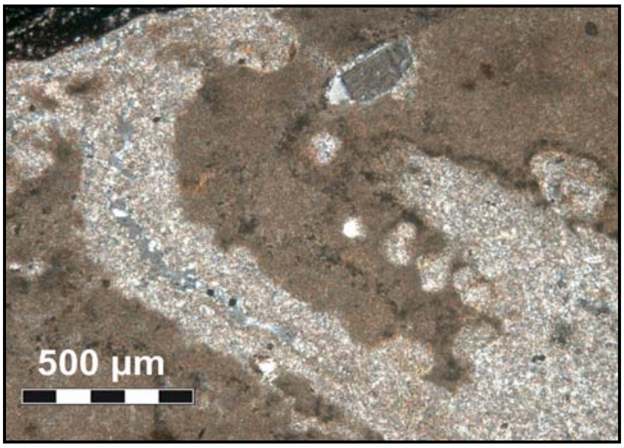
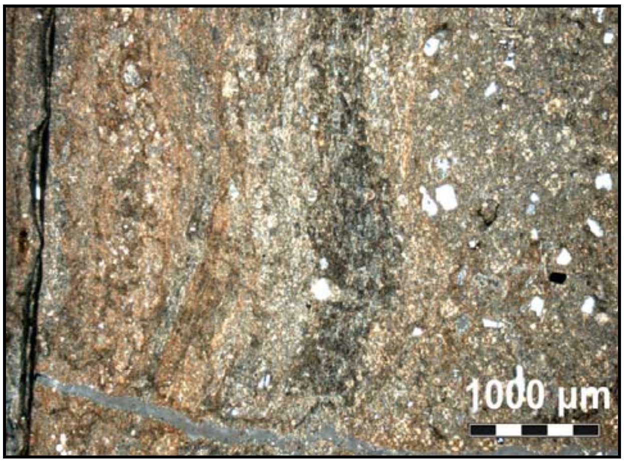
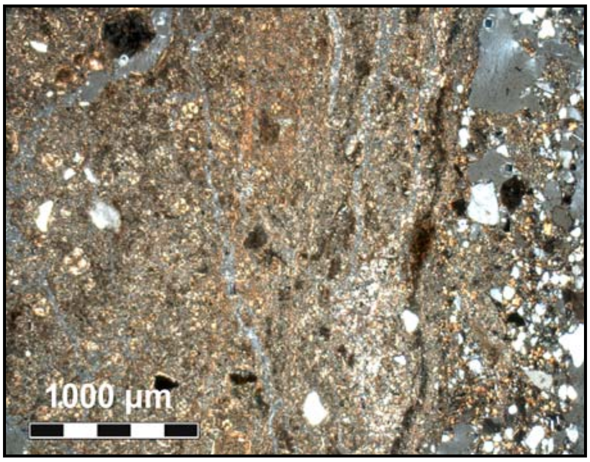
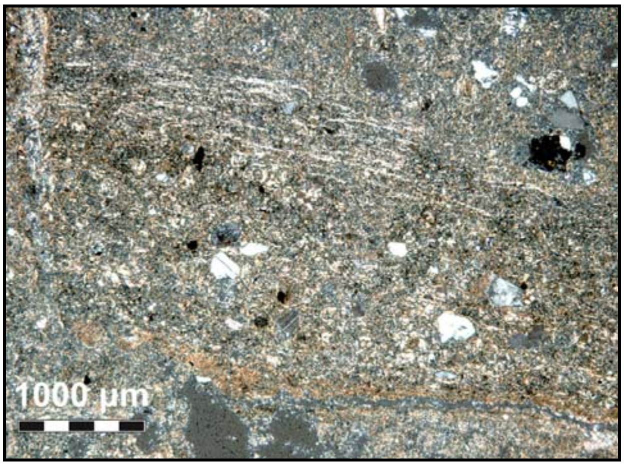
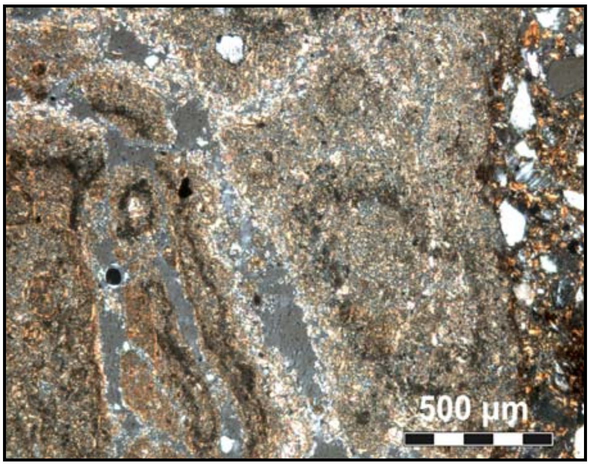


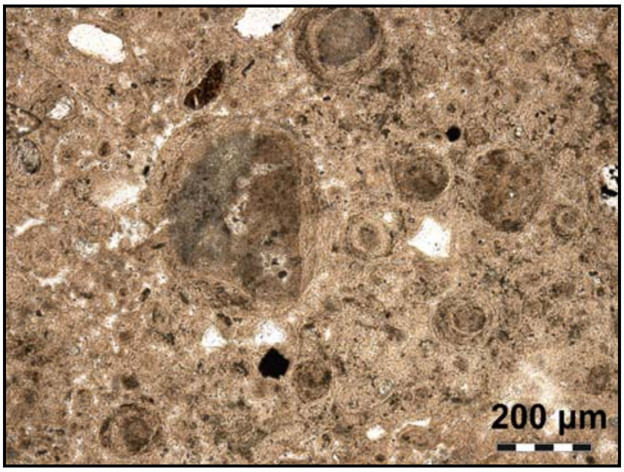
Fig. 142-144 (left to right, top to bottom): Thin sections from sample S-14 (see text for description). 142 - Note variable coloration of matrix. Whitish matrix mantles rock fragments and fills cracks (crystallaria). Small aoverall amount of clasts. 143 – Cracks within brownish matrix and rock fragments filled with whitish matrix. Note small areas of orange material. 144 – Note how crack is not completely filled with relatively coarse, whitish matrix. Crack surrounded by manganese (?) precipitates. Note the round, circular appearance of crack, pointing to solutional processes.

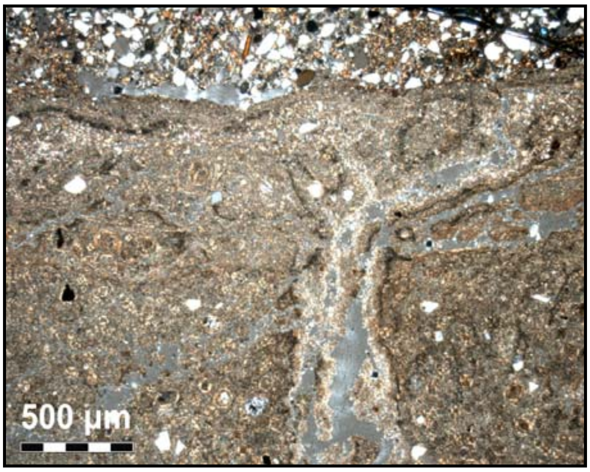












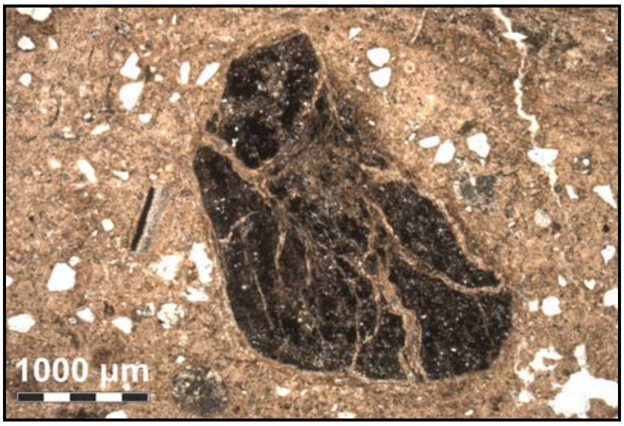
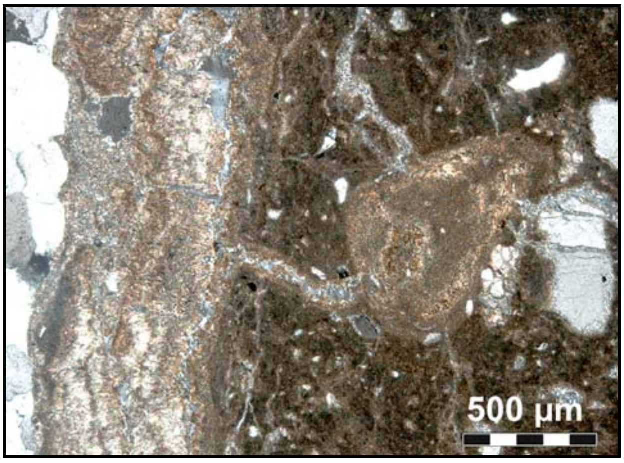
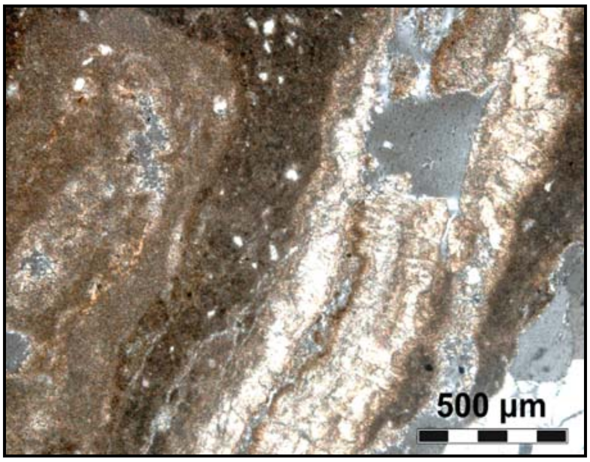
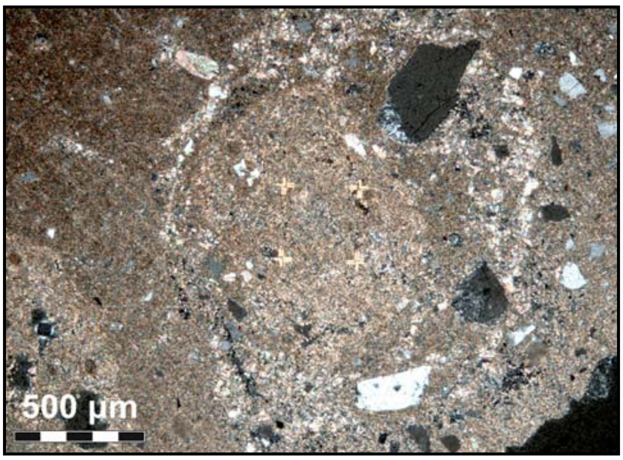
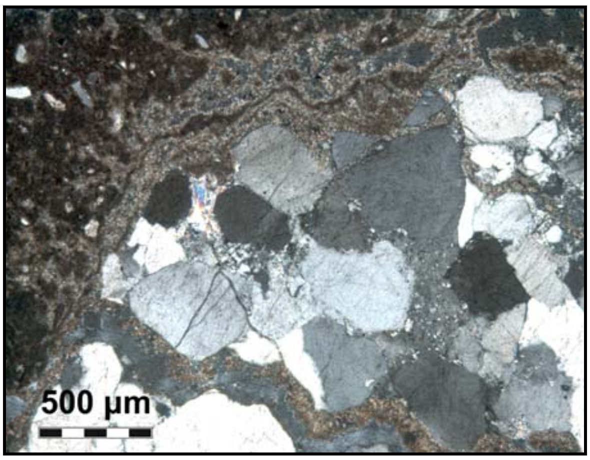


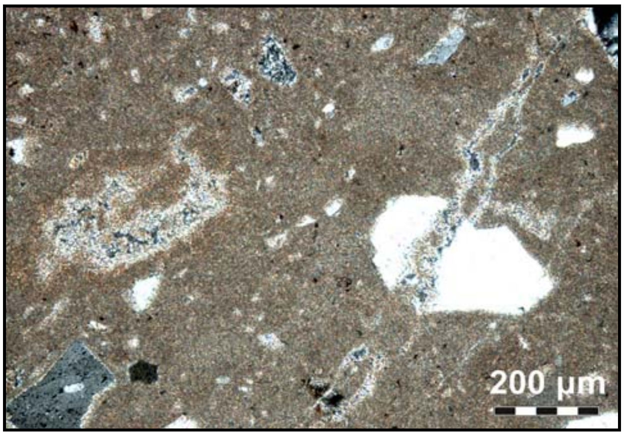
Fig. 145-151 (left to right, top to bottom): Thin sections from sample CC-14 (see text for description). 145 – Note relatively low amount of clasts. Marked lamination (colors) towards the top (left in the image). Crack in lower part through lamination. 146 – Transition from calcrete texture to sand crsut texture (right). 147 – Thin carbonate laminae. 148 – Poorly recrystallized cracks close to transition to sand crust. Note intense dark managenese precipitation and orange coloration of matrix. 149 – Abundant peloids or spheroids, partly containing matrix fragments. 150 – Fragmented overall appearance of thin section at transition to sand crust. 151 – Fragmented clast with recrystallized cracks.

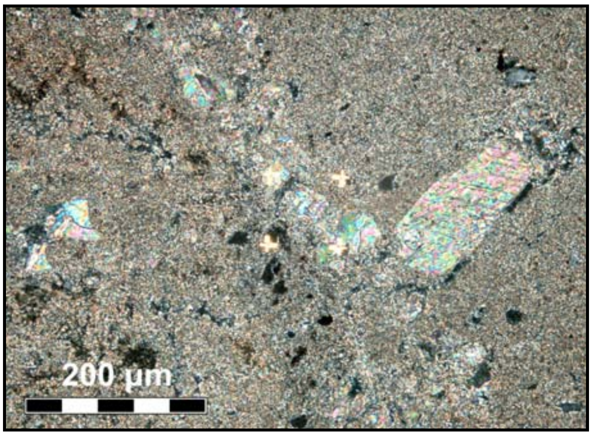












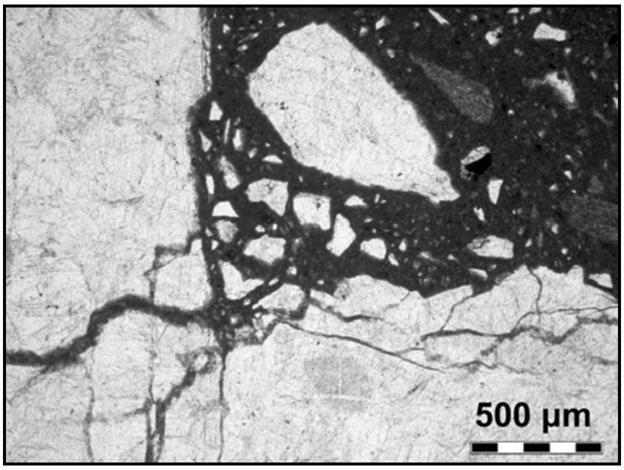
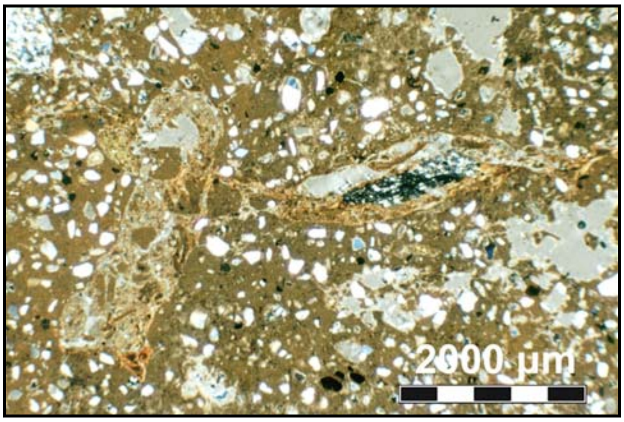
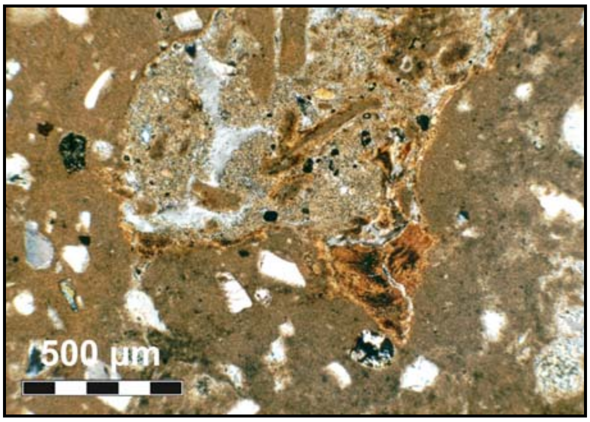
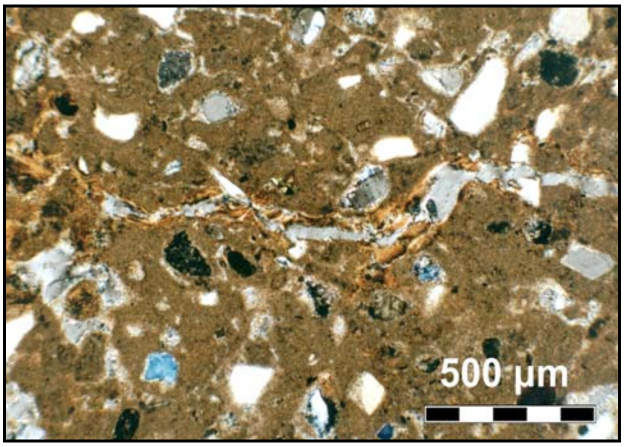


Fig. 152-158 (left to right, top to bottom): Thin sections from the sample CC-18 (see text for description). 152 – Laminar layer of carbonate around quartzitic clast (lower right). Cracks in dark matrix and fragment of lighter matrix, partly with circular orange features. 153 – Detail of carbonate laminae. 154 – Circular assemblage of matrix, clasts and minerals (glabulae?). 155 – Pallidade-like recrystallization in crack. Color changes at transition from clast to matrix. 156 – Quartz grain parted by recrystallized crack. Pore not completely recrystallized. 157 – Calcite minerals in fine matrix. 158 – Fragmentation of quartzitic clast and recrystallization by carbonate matrix.







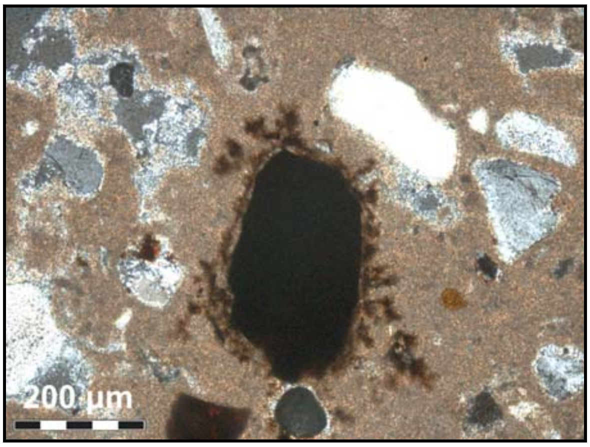


Fig. 159-162 (see preceding page, left to richt, top to bottom): Thin sections from sample CC-5 (see text for description). 159 – Calcrete of relatively dark coloration with abundant angular clasts. Many pore spaces. Note large cracks partly refilled with dark matrix fragments and light matrix (center left). 160 – Detail of 159, fill with abundant orange clay minerals (note layering), matrix and small, opaque clasts (organics?). 161 – Crack with clay coatings along walls, otherwise no recrystallization. 162 – Cemented organic material (with infiltrating acids?).

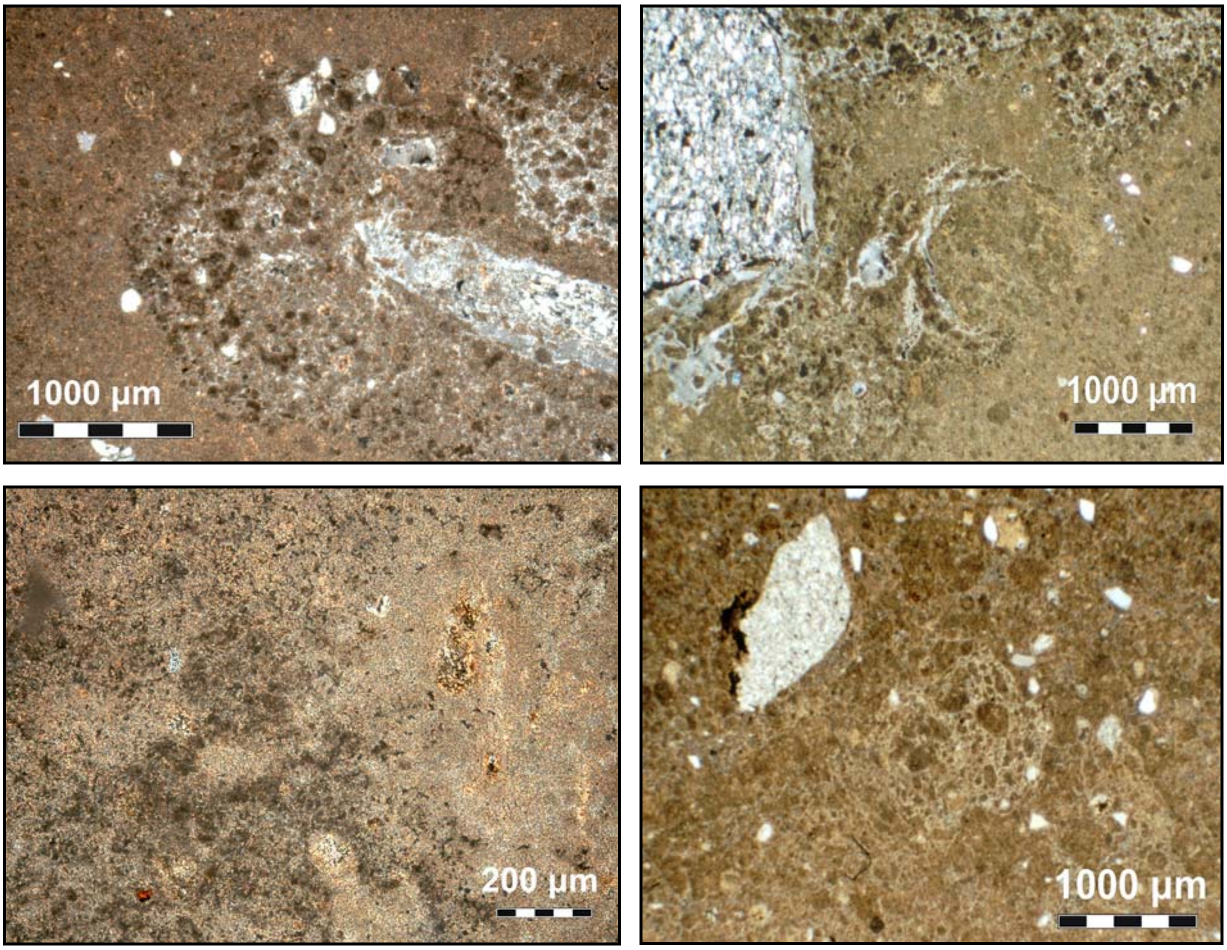


Fig. 163-166 (below, left to right, top to bottom): Thin sections from sample CC-11 (see text for description). 163 – Contrasts in matrix coloration and texture around clasts. Note abundant spheroids, peloids or nodules. 164 – Circular cracks and matrix differences around clasts. 165 – Variation in matrix coloration due to abundant, dark spots (mangenes precipitates?). Note small pore filled with orange material (clay?). 166 – Relatively large circular spheroid incorporating matrix clasts, mineral clasts and recrystallized matrix.

CC-18 This sample was collected from the rim of a drainage channel on top of the terrace T-3 at Lipán *(65,58W 23,656S)*. It appears whitish to pinkish and exhibits a rough surface, possibly signs of weathering or solution. The sample incorporates abundant clasts and appears very coarse.
The thin section shows abundant rock and clast fragments of considerable size supported by a brownish matrix. The rock fragments are quartzite, schists and phyllites. Mineral components consist of a large an poorly sorted spectrum of mainly quartz and plagioclase but also kalifeldspar, muscovite and even sparitic calcite minerals within a carbonatic matrix. The brownish matrix appears dense and cryptocrystalline, but shows several cracks and fissures. Rock fragments as well as individual minerals are fractured and cracked, while at some parts the cracks are filled with dense carbonatic matrix, at other parts, pallisade like laminated layers of cryptocrystalline matrix grow within them. Around the largest clast a thick laminar layer of carbonate has formed. In several parts the sections shows spheroidal features, partly with nucleuses of a light brownish matrix material. In most cases these features exhibit circular arrays of a very fine grained orange material, possibly illuviated clay. Sometimes these spherical features seem to be surrounded by cracks which have subsequently favoured the growth of rhombic intercalary calcite minerals.

From the above-described micromorphological characteristics several conclusions can be drawn. All of the calcretes can be classified as α-type calcretes. From the several proposed models, a model suggested by WERNER (1971) is favoured for the carbonatization and calcrete formation of the carbonated soils and calcretes within the study area (4.6.1.): decreasing $CaCO_3$- content within the profiles points to decending percolation of carbonate saturated water. Carbonate precipitation mainly results from evaporation processes within the soil, the biogenic component is less important. Depending on local morphology, lateral movements of water, either by surficial overland flow or by interflow, may transport and precipitate carbonate as well.

The *morphological setting* of the calcretes discussed above seems to be reflected by different micromorphological characteristics. Calcretes at settings with dominant overland flow due to pronounced morphology (e.g. high slope angles) have abundant pisolithic features, peloides and spherical nodules of reworked calcrete material. In contrast, calcretes at scarps or rims (e.g. along the terrace top) seem to show a high amount of floating clasts. In addition, the matrix is relatively dark. This points to interflow processes as the main agents of carbonatization. The interflow water gets saturated not only with $CaCO_3$ but also transports organic material and clay through the upper soil horizon and precipitates the material as soon as it exits the solum, in this case at vertical rims and scarps of the terraces. The rather porous character of the samples CC-5 supports this assumption.
Most of the samples bear crystallaria, cracks and fissures filled with secondary carbonate matrix. This might point to the *multiphase character* of the calcretes due to alternating phases of calcrete formation and calcrete desintegration coupled with clay illuviation and manganese precipitation. These alternating phases have been interpreted to reflect various climat changes, possibly oscillating between semi-arid and semi-humid.
The dense and homogenous, but nevertheless polygenetic character of the matrix in sample S14 might indicate a *mature stage of calcrete formation*, presumably corresponding to its morphological location on top of the older T-2 terrace. Particularly the rounded shape of many pores and cracks might point to slight karstification, very likely under much wetter

conditions than today. A clear laminar layer, typical for a mature calcrete, has exclusively been observed within the sample CC-14. Very likely this might be attributed to the samples morphological situation below a topographic step again. Due to the enhanced water supply a faster and laminar growth of the calcrete might as well have been the result of microbiogenic activity as postulated by VERRECCHIA (1994).

The question for the source of the carbonate material cannot be answered on the base of the available data. A provenance from local carbonate rocks seems unlikely due to the very limited outcrops of limestone within the study area. Possibly rock weathering contributes some calcium to carbonatization supported by airborne carbonatic dust typical for semi-arid to arid regions. Neither macro- nor micromorphological analysis have contributed further information regarding this problem. In any case, *"much work is needed before the links between calcrete morphology/micromorphology and climate/biology are clear enough for their use in paleoenvironmental interpretation"* (WRIGHT AND TUCKER 1991, p.7).

4.6.3. SAND CRUSTS

The term „*sand crust*" refers to a very characteristic horizon of cemented sandy material frequently observed in the study area. Particularly its hardened appearance and morphological resistivity against erosional processes (Fig. 168 and 170) justify the denomination as "*crust*". The sand crust has been observed within soil profiles as well as at the surface, where it seems to occur areally, predominantly in the upper study area (Potrero Grande, Quebradas de Lipán, Sepulturas and Estancia Grande). Due to fluvial incision, it appears very pronounced in the vicinity of drainage channels where it is always inclined towards the channel (Fig. 167). In this way it traces a former landscape surface. Two thin sections of sand crust samples have been analysed to obtain further information regarding the processes which have formed the crust and the material building it up.

Fig. 167: Sand crust on top of T-3 terrace surface at Lipán, clearly inclined towards the drainage channel and subsequently exposed due to channel incision.

Fig. 168: Sand crust at Potrero Grande. Note the desintegrated appearance of the crust and the overlying clast (top right) implying local burial of the crust before exposure due to erosional processes.

Fig. 169: Sand crust of several meters thickness at Potrero Grande, subsequently dissected and eroded.

Fig. 170: Sand crust on top of T-3 terrace surface at Lipán. Sand crust has been buried by brownish fanglomerates and only recently been exposed by gullying.

CS-17 This sample was collected from the surface of an A-1 alluvial fan in the Quebrada de Sepulturas (*65,591W 23,651S*). Its macroscopic textural characteristics are similar to a sandstone with a orange sandy matrix supporting a few larger clasts. Several lichens grow on the surface.

The thin section shows a clast supported fabric with most grains being fine sand. The sorting is extraordinary well and only a few larger clasts seem to float in the otherwise uniformly distributed fine sand. These clasts and rock fragments are slightly rounded and seem to be of quartzitic and phyllitic origin. Most grains are subangular to subrounded. Quartz and plagioclase are the most frequent components, while kalifeldspars, heavy minerals, epidote, hornblendes and opaque grains occur to a much lesser extent.

The sample shows many pore spaces. In between the grains a cryptocrystalline orange matrix has established, possibly being hematite-rich illuviated clay minerals. Locally these clay minerals have reached remarkable sizes.

CS-18 This sample was collected from the rim of a drainage channel on top of the terrace T-3 at Lipán (*65,58W 23,656S*). Its macroscopic textural characteristics are similar to CS-17 but its surface shows lichen growth of up to several centimeters of size.

The thin section shows a clast supported fabric. Most grains are subangular to subrounded and the grain size ranges between fine and middle sand. Several subangular larger clasts and rock fragments, mainly phyllites, quartzites ans sandstones, are supported by the sand mass. Mineral grains are dominantly quartz, some plagioclase and to a lesser extent kalifeldspar, horblende and heavy minerals.

The sample shows many larger pore spaces. Its grain size composition is slightly coarser than sample CS-17 and sorting is not as good. Overall fabrics are very clast-supported and very little cryptocrystalline orange matrix material can be observed between the clasts, possibly resulting from the illuviation of hematite-rich clay material.

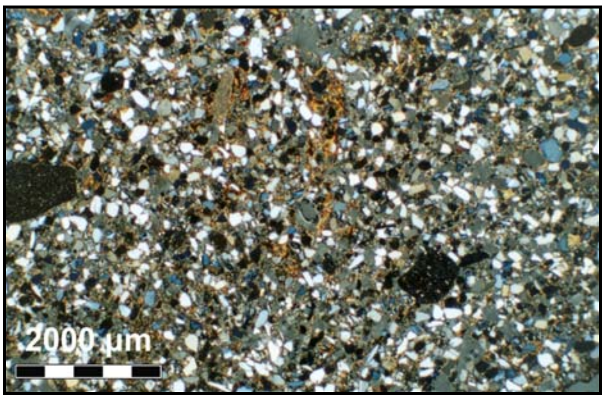


Fig. 171: Grain-supported fabrix of sample CS-17. Note the good sorting, frequent pores spaces (grey) and the orange matrix in between grains.

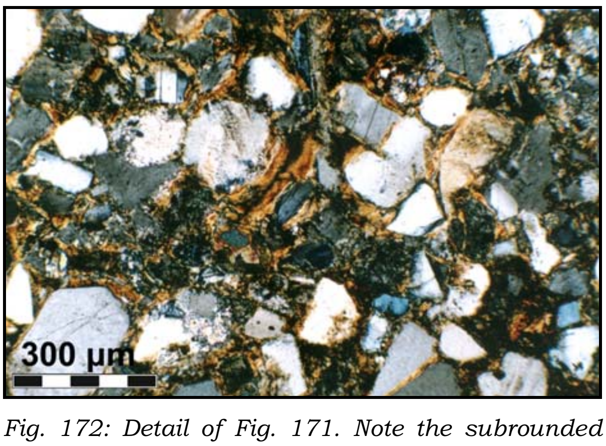


Fig. 172: Detail of Fig. 171. Note the subrounded character of the mineral grains and the relatively large sized clay minerals (orange).

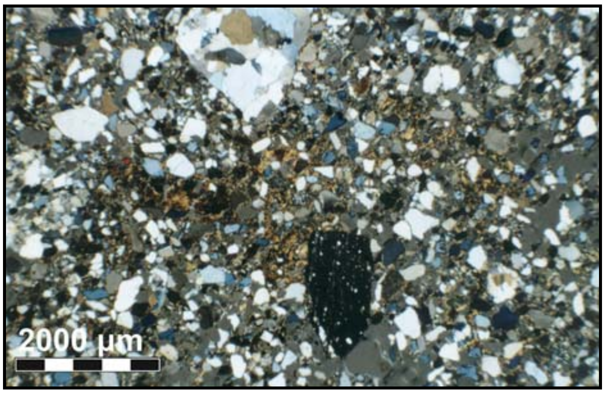


Fig. 173: Grain-supported fabrics of sample CS-18. Note large pore spaces (grey) and the poor quality of sorting compared to Fig. 171.

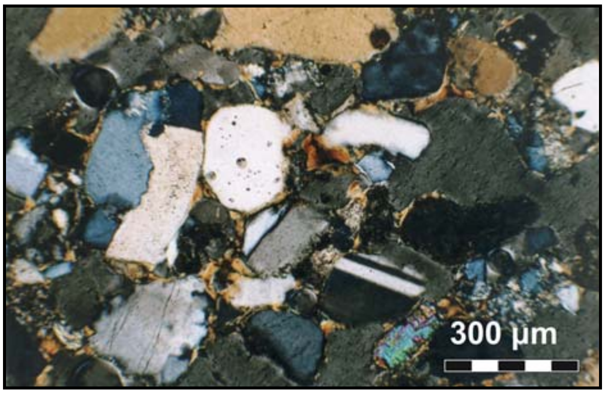


Fig. 174: Detail of Fig. 173. Note subrounded grains of the middle sand fraction, large pore spaces and the little amount of orange matrix.

From the above described macromorphological and micromorphological characteristics several conclusions can be drawn. First of all, the sand crust does not show any sign of carbonate cementation. Its cementation results exclusively from the crystallization of relatively little amounts of clay minerals in the inter-grain pore spaces. Pore spaces are abundant between the grains of fine and medium sand. Therefore it is remarkably that the crust has a high erosional resistivity. Under humid climatic conditions, the clay cement would have been dissolved much easier. This would have led to the desintegration of the crust. Thus the well-preserved crust most probably implicates that since its deposition and cementation climatic conditions have never been significantly wet again.

The high amounts of feldspars possibly point to very reduced intensities of chemical weathering, while the source material of these minerals can not be determined. The predominance of fine and medium sand might point to a eolian component in transport, while the larger floating grains and rock fragments imply at least some fluvial reworking for the further transport and final deposition. Eolian transport and deposition is indicated by the thick accumulations of sand crusts in positions of wind shadow (e.g. below the rims of older terraces, in vicinity of former drainage channels etc.). However, the source area of the eolian material must have been within a relatively short distance in the order of tens of kilometers as many grains are not ideally rounded.

Summing up, the sand crust could be interpreted as a fluvio-eolian sediment, subsequently hardened by the illuviation and drying of clay material which has most likely been contributed by the reworking of cambisols or resulted from increasingly pronounced processes of chemical weathering.

4.6.4. PRELIMINARY RESULTS

As a conclusion from the pedological observations described above, a sequence of soil formation processes can be established. Significant soils have only been observed on top of terraces and alluvial fans. Everywhere else, morphological and environmental conditions have not been stable enough.

The oldest phase of soil formation is resembled by a clay-rich reddish horizon described as cambisol which underlies all other horizons in most cases. This soil horizon has been observed on the terraces T-1 and T-3, while on terrace T-2 a multiphase calcrete has been observed. These two soil types do not seem to be genetically related to each other, as they have not been observed in close association to each other in any soil profile. Instead several types of calcic soils and calcretes seem to be overlying the cambisols, frequently seperated by a remarkable layer of indurated sand which even shows up in present day morphology. However, several, at least two phases of carbonate accumulation and calcrete formation interrupted by phases of calcrete desintegration and enhanced clay illuviation can be inferred from macromorphological and micromorphological observations.
Paleoenvironmental interpretation of the various phases of soil formation is difficult due to the very localized pedological data from the study area and limited knowledge regarding the complex connection between environmental conditions and pedogenic processes. Tentatively, the formation of cambisols and luvisols would indicate slightly subhumid conditions while carbonatization and calcrete formation points to semi-arid to arid conditions. Formation of the sand crust demands a lowered rate of chemical weathering coupled with reduced vegetation due to eolian processes involved, which should imply reduced temperatures and precipitation.

4.7. FLOODPLAIN MORPHOLOGY

In contrast to the extended slope areas, comparetively little space of the study area is taken up by the valley floors. The transition between the slope and the valley floor is in most cases very marked and narrow, typically characterized by adjacent active alluvial fans, badlands and steep slopes in bare rock or even vertical walls. Only at very few places within the study area, the valley floor is broad enough to allow a transition to the adjacent slopes stable enough for the establishment of settlements like the village of Purmamarca and Patacal.

Otherwise the valley floor is entirely buildt up by a floodplain. At least in its lower reaches the longitudinal gradient of the flooplain is about 2° and increases to 3° to 4° close to Lipán. The width of the floodplains varies significantly, depending on the overall morphological setting, ranging from a few tens of meters up to 300 meters. Particularly in the upper reaches of the study area as well as in the smaller quebradas the floodplain may not exceed several meters of width. Within the floodplain, parts of higher activity can be distinguished from less active or presently inactive parts by the amount of vegetation growth. Within the floodplain, these active parts are essentially free of vegetation. They seem to show a very weak tendency to meander from one side of the floodplain to the other. The wavelenghts of these oscillations range from several tens to several hundreds of meters. Therefore the inactive and plant-covered floodplain parts seem to be located preferentially on the inner floodplain banks.
While the cross profile of the floodplain has been observed to show a slight overall tendency to convexity, the channel patterns seem to vary enormously from quebrada to quebrada. At some places, braided channel systems have developed indicating a low resistitvity of the floodpain deposits. In contrast, the stream channels are remarkably straight at other places, pointing to high resistivity of the floodplain deposits and low transport capacities of the streams.

Generally speaking, discharge and sediment load are the two controlling variables of fluvial floodplain morphology. They determine the stream's capacity to erode, transport and accumulate material (KNIGHTON 1998). In other words, differences between the channel patterns can theoretically be attributed to hydrological or sedimentological controls. However, the contrasting stream channel patterns have been observed with no apparent regional trend. Sometimes channel behavior even seems to shift between straight and braided several times within the same quebrada. Therefore sedimentological differences within the floodplain rather than significant differences in runoff seem to control the local stream channel pattern. The sedimentological composition of the floodplain has been observed to be subject to enormous variations. All transitions between clay and silt-sized material and boulders of up to 80 centimeters have been found on the active floodplain.

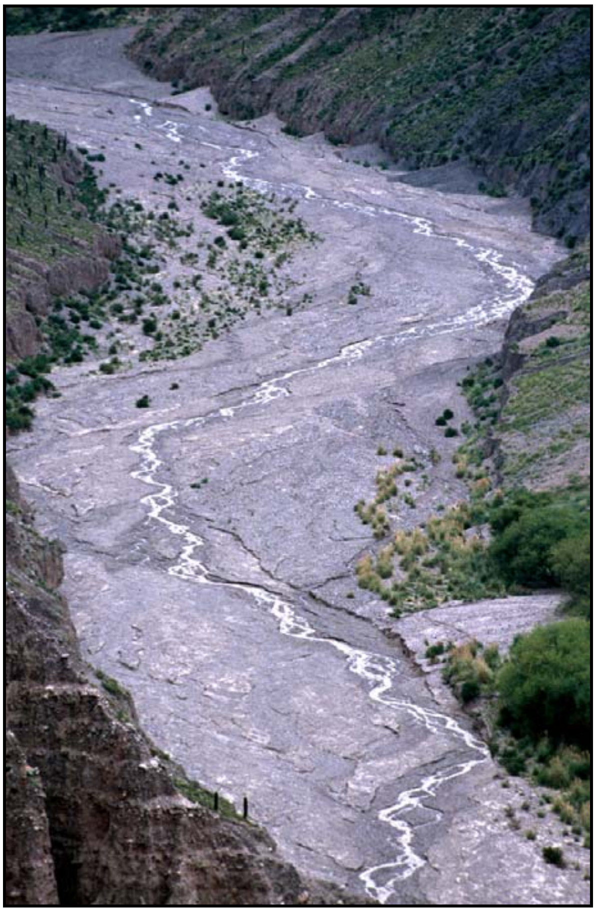


Fig. 175: Floodplain morphology in the Quebrada de Lipán. Note the sediment supply by lateral alluvial fans, active and inactive floodplain parts, and tendencies to meandering (medium-size) and braiding (small-size).

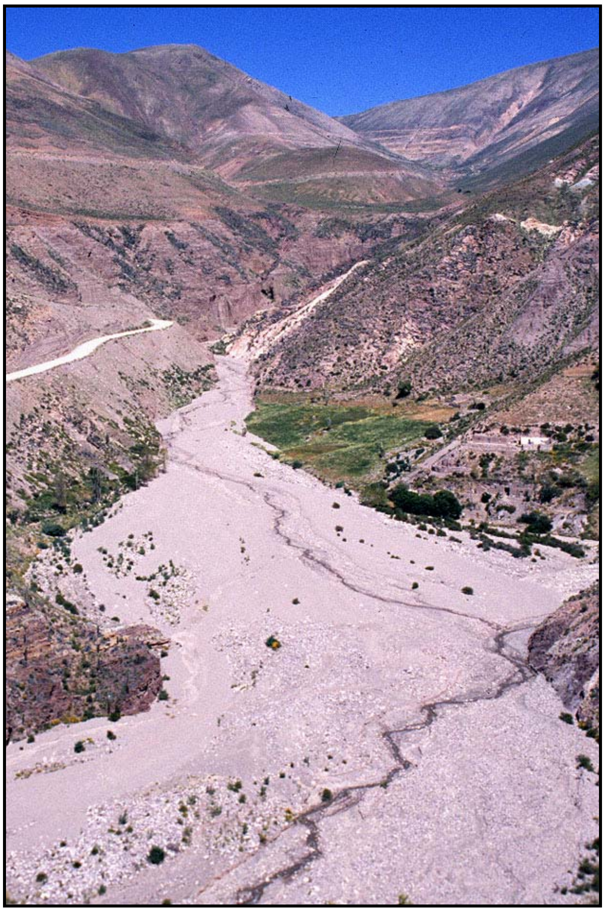


Fig. 176: Floodplain morphology in the Quebrada de Potrerillos dominated by debris-flows. Note fan-like shape of floodplain.

The major part of the clasts consists of quartzites, but also schists and phyllites. Corresponding to the geological situation of the study area, several other rock types like andesites, sandstones, shales and limestones have been identified within the floodplain sediments further downstream, but make up only for a small percentage of the entire deposit. While most clasts are subangular, particularly the andesitics and sandstone clasts are remarkably rounded. This points to their relatively low erosional resistivity, confirmed by PAREDES ET AL. (1998). Even though relatively large clasts of more than 15 to 20 centimeters are common in all quebradas, the sorting of grain sizes shows significant differences.

Particularly in floodplain segments where braiding has been observed to be the dominant channel pattern (Fig. 175 and 177), very few clasts larger than five to ten centimeters have been found, indicating a relatively good sorting by fluvial processes. Most of the finer grain sizes concentrate in the lateral parts of the floodplain or have been entirely removed. The morphology of the floodplain surface is comparetively regular and flat. Due to the frequent shifts of the braided channel system, lateral erosion and undercutting of adjacent slope areas has been noted.

Fig. 177: Fluvially dominated floodplain (Quebrada de Lipán) with a temporary braided channel system. Note smoth overall aspect of floodplain surface.

Fig. 178: Floodplain controlled by sediment input from debris-flow depostion (upper Quebrada de Purmamarca); note fluvial erosion (left) and surface roughness of debris-flow deposit (foreground).

In contrast, floodplain segments with straight channel patterns (Fig. 176 and 178) show a very poor sorting. High amounts of fine grain sizes mixed with coarser clasts up to 50 centimeters are found to form a pronounced surface morphology. In fact, these sediments have been recognized as debris-flow deposits due to a variety of characteristics. The viscous mode of flow can be inferred from the way they onlap obstacles (Fig. 179). Lobe-like deposition points to their high density (Fig. 180). The relatively high quantity of fine grain sizes in the matrix of these debris-flow deposits manifests itself in the amount of sediment consolidation, also leading to dessication cracks (Fig. 181). In most cases the debris-flow deposits have partly been eroded by subsequent discharge, usually leaving behind the coarser clasts. Thus, the quantity of sediment supplied to the floodplain by debris-flows exceeds the potential of the stream's transport capacity. Consequently, the stream is confined to a very thin and narrow channel, and floodplain morphology is largely controlled by debris-flow deposition.

Fig. 179: Fresh (2001) debris-flow deposit close to Lipán. Note how little harm viscous the 50 cm flow has done to the bush.

Fig. 180: Small lobe of debris-flow (2001) spilled over earlier deposit indicating high flow viscosities.

Fig. 181: Deposit from the 2001 debris-flow on the floodplain close to Lipán (foreground) being eroded by fluvial action.

Debris-flow deposits on the floodplain have been found within in the entire study area. However, their morphological appearance and extent varies from quebrada to quebrada. While in the Quebrada de Purmamarca between the Potrero Grande and La Ciénaga a single debris-flow deposit on the floodplain could be traced for more than five kilometers downstream, debris-flow deposits in the Quebrada de Lipán concentrate on small lateral alluvial fans, classified as A-3 fans (4.5.). Corresponding to this observation, floodplain morphology of the two quebradas varies significantly. While the Quebrada de Lipán shows a tendency to braiding, the upper Quebrada de Purmamarca has irregular and straight stream channel patterns. Judging from the overall drainage patterns and catchment sizes, this difference in debris-flow input may be linked to the sizes of the catchment areas of each quebrada. Larger catchments automatically gather larger amounts of runoff and water, increasing the flow capacity of the debris-flow. Therefore sediment input by debris-flow deposition is much more important for the floodplains in the larger quebradas.

Larger precipitation events result in an immediate increase in discharge rates and transport capacity of the stream due to higher flow velocities and turbulence (Fig. 182). Only during these low-frequency events the stream has the capacity to erode, transport and modify floodplain morphology. Aside from the transport of larger clasts, the streams appear to transport and deposit remarkable amounts of suspended load of fine grain sizes. Badland areas as well as the debris-flow deposits within the floodplain present sufficient source areas for these sands, silts and clays. Commonly, this loamy sediment gets deposited in relatively large sheets or lenses of fine material in the lateral parts of the floodplain where discharge and turbulence are smallest (Fig. 183). Where this material is not buried by subsequent flooding or debris-flow deposition, it is prone to eolian transport after drying up.

Fig. 182: Increased discharge after precipitation event close to Purmamarca (04-2001). Note turbulent water surface indicating high discharge velocities and water color due to a high quantity of suspended load.

Fig. 183: Lenses of loamy material resulting from the depostion of suspended load in lateral part of the floodplain. Location directly adjacent to Fig. 182.

From these observations it becomes obvious that the present floodplain is subject to a variety of processes. Debris-flows contribute material to the floodplain. This occurs either by lateral input from small alluvial fans or by deposition of larger debris-flow events which may affect the floodplain within a distance of several kilometers. In addition, fluvial processes rework and distribute material within the floodplain and remove it from the drainage basin. The present floodplain is therefore subject to a variety of remarkable short-term modifications. Within less than two years a dirt road at Lipán has been observed to disappear completely due to intense aggradational processes and plant growth (Fig. 184 and 185). Many further examples for the present morphodynamic situation of the floodplain have been discovered at various places within the study area, most of which show the burial of plants and trees by sediment aggradation (Fig. 186).

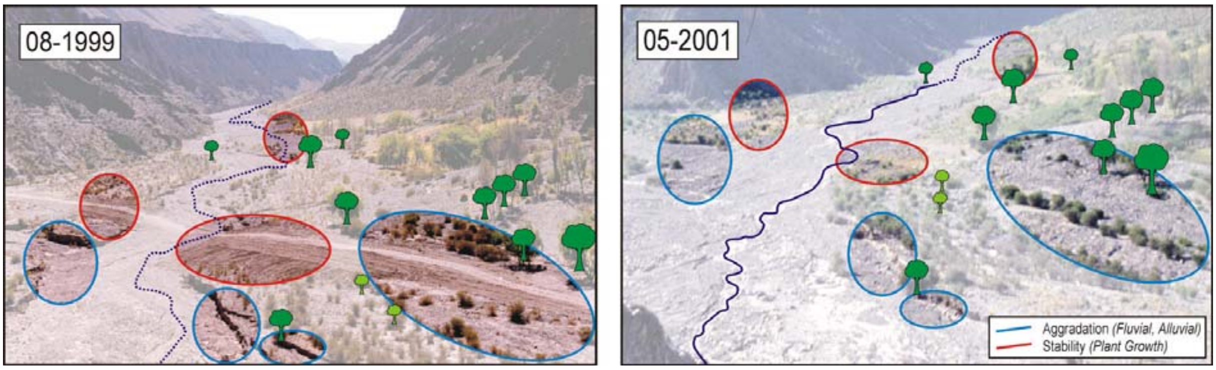


Fig. 184 and 185: Examples of present fluvial dynamics and floodplain modifications in the Quebrada de Purmamarca close to Lipán. Note the tendency to aggradation in the river bed and aggradation by debris-flow / alluvial fan activity (large blue circle); aggradational tendency does not seem to be restricted to the valley floor.

Summing up these observations, it seems that reworking, transport and removal of the sediment by fluvial processes do not keep pace with the rate of debris-flow deposition at present. This enormous input of sediment corresponds to the observations of intense gullying, slope dissection, badland formation and intense deposition of the A-3 alluvial fans mentioned above. All these are potential sources of debris-flow material. However, the recognition of this significant tendency of floodplain aggradation on the base of observations within the study area has also been confirmed by locals and several authors (AGUERO 1986, CHAYLE AND WAYNE 1995, SEGEMAR-ITGE 1998, SOLER AND MAY 2000).

The tendency of floodplain aggradation has very direct effects on the people living in the study area. Due to the high altitudinal differences and the steep slope morphology, the only possible setting for settlements, agriculture and infrastructre is the floodplain and the lowermost slopes. Therefore, the enormous morphodynamic activity of these settings constitutes a severe danger for the people. In this context, particularly debris-flows have been shown to be an integrative part of the morpodynamic system of the these settings. While the smaller events mainly contribute to the slow local and regional aggradation of the floodplain and do harm only to very limited areas, the larger low-frequency high-magnitude events are capable of threatening many lives as well as causing high economic damage. This situation of enormous morphodynamics typical for semi-arid mountain areas has always constituted a burden for regional development, not only within the study area but also in the entire region of the Quebrada de Humahuaca. Therefore several authors have commented on the demand for additional means in order to prevent and protect the people from these natural events (AGUERO 1986, CHAYLE AND WAYNE 1995).

Fig. 186: Floodplain in the Quebrada de Sepulturas covered entirely with debris-flow deposits. Note the buried "precaution" sign indicating the course of a gas pipeline.

Fig. 187: Building of the former railway station of Purmamarca in the Quebrada de Humahumaca which was buried by a single debris-flow event in 1984.

Within the study area, first actions have been undertaken to prevent damage by aggradation and debris-flows. These actions have so far only focused on very few places in close relation to settlements like Purmamarca, Patacal and Lipán. Mainly, the construction of artificial channels (Fig. 190) and wooden obstacles (Fig. 188 and 189) is supposed to prevent debris-flows from running over fields or populated areas. The efficiency of wooden fences to divert debris-flows is at least questionable against the background of Figure 189 where a single debris-flow event has buried almost the entire fence of two meters hight. From the village of Purmamarca, SOLER (2002) has reported the floodplain to have been elevated several meters above the present village level. In order to prevent overflowing, flooding and burial by debris-flows, locals have buildt artificial walls and levees along the floodplain (Fig. 191).

The question whether this wall will also resist a high-magnitude debris-flow event may be of vital importance for the entire municipality of Purmamarca. However, all these anthropogenic activities can be regarded to be the ultimate and most recent morphological modifications within the floodplain and are therefore a small but essential component in the landscape evolution in the Quebrada de Purmamarca. The above described observations lead to a number of preliminary considerations.

Fig. 188: Wooden fence on the floodplain at Purmamarca to divert debris-flows and flooding and prevent overflowing into village.

Fig. 189: Wooden fence of similar construction as Fig. 188 at Lipán. Note amount of burial the fence has experienced by a single debris-flow event in 1999.

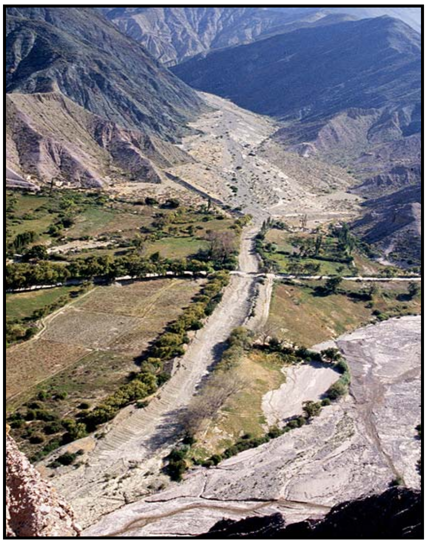


Fig. 190: Artificial drainage channel at the Quebrada de Coqueña to prevent flooding and burial of agricultural land.

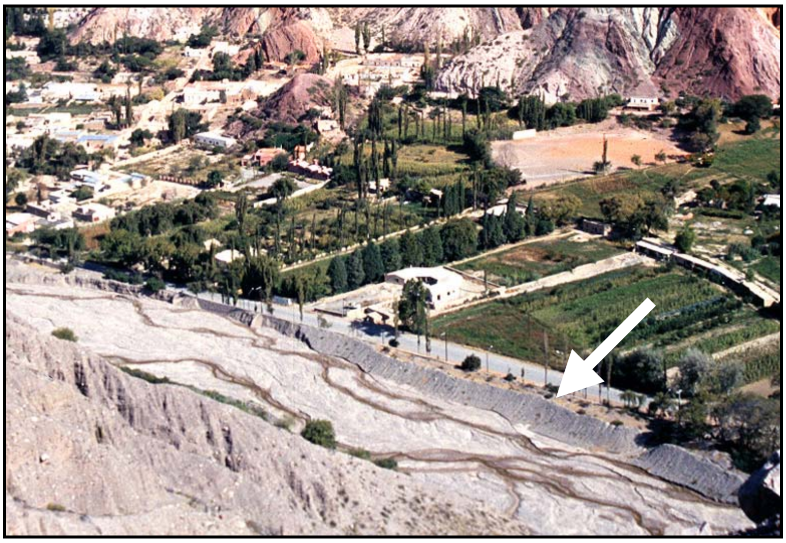


Fig. 191: Artifical levee (2-3 meters high) along the floodplain at the village of Purmamarca (background) to prevent overflowing into village.

From the broad, depositional character of the floodplain it seems likely that its present overall morphology is the result of enormous accumulations of sediment. In other words, the valley floor is supposed to consist of a thick fill of alluvial and fluvial deposits of unknown depth. The deposition of these sediments must have commenced after the intense phase of incision which has postdated the deposition of terrace T-3 and A-1 alluvial fans. This assumption is corroborated by the partial burial of the landslide deposits at present floodplain level.

However, the present floodplain activity can be divided into two representative types. Where sediment supply is low and sufficient discharge is available, the floodplain is dominated by fluvial processes and tends to develop a braided channel system. In constrast, where sediment supply exceeds the stream's transport capacity, fluvial processes loose importance and the irregular floodplain morphology is controlled by sediment deposition. In this context mainly debris-flows are responsible for the extraordinary large quantity of sediment supply. In accordance with additional observations mentioned above, the debris-flows receive the bulk of their material from the ongoing dissection of most slope areas.

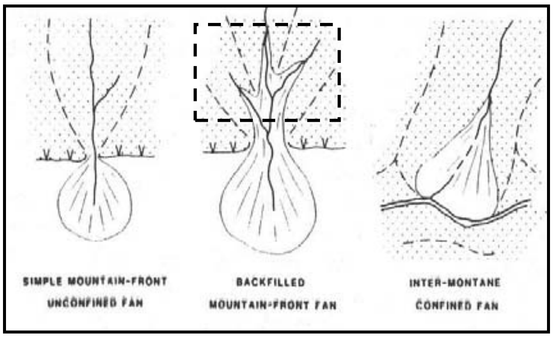


Fig. 192: Geomorphological setting of floodplain in the study area with dominant alluvial processes (from Harvey 1997, p.232).

Therefore, due to this dominance of alluvial processes and judging from the morphological and sedimentological properties of the floodplain, great parts of the valley floor in the study area resemble an alluvial fan with pronounced confinement to the valley dimensions much rather than a fluvially dominated floodplain (Fig. 192). Investigations in the Tibesti Mountains in Africa by OBENAUF (1973) confirm these observations and show that alluvial processes and forms are integrative components of dryland fluvial systems.

The observations of floodplain morphology and dominant processes seem to confirm an aggradational trend which has been reported for the entire region by a variety of authors. This trend manifests itself at several places where depositional processes on and along the floodplain have buried and damaged plants, fields, roads and even buildings. Consequently, a variety of artifical modifications of the floodplain have been implemented in order to prevent further damage.

4.8. EOLIAN PROCESSES

Even though the semi-arid conditions in combination with the relatively loose vegetational cover should allow a significant mobility of fine material with grain sizes of sand to clay, only very few observations point to important eolian processes in the study area.
No major landforms resulting from sand accumulation like dunes have been detected during field observation and analysis of remote sensing data. Even where larger accumulations of fine material have been found, e.g. in the floodplain, indications for eolian processes like ripple marks were absent.
Nevertheless, as mentioned above, pedological investigations have shown some evidence for past eolian activity (4.6.3.). In particular the sand crusts observed on the terrace surfaces and A-3 alluvial fans have revealed an extraordinary good sorting. Grain size composition of medium to fine sand has been interpreted to be typical for eolian deposition. Most grains have been found to be subrounded, and to a lesser extent subangular and rounded. This would imply relatively short distances of transport and has been observed from many eolian deposits. However, the sand crusts contain small percentages of clasts sizes too large for any eolian transport. Thus fluvial reworking must have occured before deposition, if the good sorting can be attributed to eolian activity.
An additional geomorphological observation supports the interpretation of these sand crusts as fluvio-eolian deposits. An increased thickness of sand crusts deposits has been observed below the eastern and northern terrace scarps at Potrero Grande (Fig. 168 and 169), on top of the terrace surface at Lipán at the eastern foot of the slope and in association to slight depressions along pre-existing drainage channels. Therefore it has been concluded that sand crusts have been found predominantly in positions where a combination of enhanced slope wash and wind shadow was given. As wind directions are supposed to have been stable at least for most of the Pleistocene (GREENE 1995), dominant wind directions in the study area should have been from the southwest (RUTHSATZ 1977). Consequently, sand has been trapped on the eastern and northern slopes and in depressions like drainage channels, and has subsequently been washed and accumulated before it dried up and hardened.

Another interesting landform of larger size has been observed on the deeply dissected terrace surface of terrace T-1 at the foot of a slope with eastern aspect at Potrero Grande (Fig. 193). Although no detailed field data is available, a ramp-like deposit of approximately 300 meters width extends several tens of meters upslope. Its slope profile is concave and the slope angle is everywhere higher than the terrace surface below and increases upslope. In contrast to the surrounding terrace, the deposit has a very light brownish color and seems to consist of finer grain sizes than the terrace deposit. It has been subject to intense dissection. Without further knowledge no definite conclusions can be drawn, but due its setting below a slope of east aspect it presumably consists of allochthonous eolian material.

The obvious fluvial dissection of this feature implies a phase of consolidation postdating its deposition. Conclucing from these characteristics this landforms might be interpreted as a relict fluvio-eolian deposit, possibly corresponding to a sand ramp, formed by similar processes as the above discussed sand crusts.

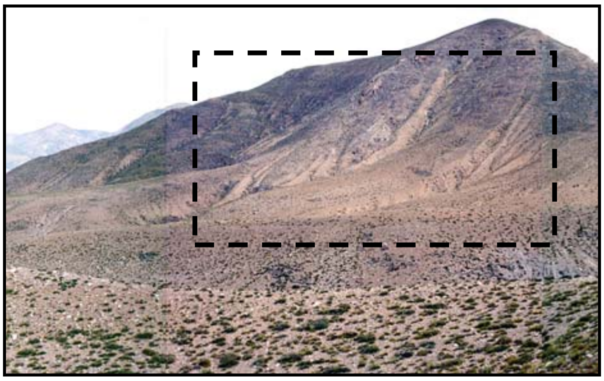


Fig. 193: Ramp-like feature on the top of the terrace T-1 at Potrero Grande (box). Note the contrasting light brownish color of the deposit.

Despite this lack of more evident eolian forms, wind activity is reported to be of great climatic importance anywhere in the Cordillera Oriental. This is particularly true for the larger valleys connecting lower areas with the Puna plateau, where wind-free days are practically absent (RUTHSATZ 1977). Due to insolation differences wind is usually blowing uphill during the day before changing to downhill directions during the night (ENDLICHER 1995). However, wind intensity during the night has not been noticed to be of the expected intensity. This phenomenon has been confirmed by RUTHSATZ (1977) and might be explained by an exchange process dominantly active in higher air masses. Therefore, the relatively low intensity as well as the frequent changes of wind direction might be responsible for the lack of present eolian forms. In addition the existence of desert pavements on the extended terrace surfaces as well as on some slope areas prevents any mobility of finer grain sizes. The sealing effect of the large amount of pebbles at the surface protects these pavements and prevents the outblowing of fine grain sizes.

However, at several places, particularly in the lower study area (e.g. Quebrada de Tumbaya, Quebrada de Sunchoguaico) desert pavement has been observed to be underlain by a horizon of loose, fine-grained, powder-like material of carbonatic composition. This material shows relatively poor sorting and contains several larger clasts of up to five centimeters. $CaCO_3$-content ranges between 15 and 35 % and mm-sized $CaCO_3$-nodules are common. In addition carbonate coatings not thicker than one millimieter mantle most of the clasts. Otherwise the horizon is entirely unconsolidated and does not show any remarkable structure. The investigation in the field has not been able to estimate the depth of this horizon.

As described above, the existence of a desert pavement on the terrace and alluvial fan surfaces is interpreted as evidence for the inactive character of the surface. The pebble concentration protects the underlying fine sediments from erosion and deflation and prevents significant infiltration of surface wash. Several mechanisms for the formation of desert pavement have been proposed, the most commonly known being deflation, water

sorting and upward migration by cycles of freezing and thawing or wetting and drying respectively (COOKE ET AL. 1993).
Particularly the larger pebbles and cobbles on the pavement surface have not been observed to be supported by fine-grained material. Instead they are exposed at the surface, resting on sand and pebbles. This might indicate some importance of eolian deflation for the removalt of the fine material from the surface. On the other hand, doubts have been voiced as to what extent wind is capable of removing sand and silt from the otherwise stable and consolidated surface (COOKE ET AL. 1993). This is particularly true for the study area where not even a minor eolian accumulation has been observed. In addition, deflation cannot account for the observed accumulation of fine material below the surface and deflation is not likely to be the only responsible component of desert pavement formation. Where sorting and particle concentration is reported to be the result of wetting and drying cycles, subpavement soil is typically characterized by a vesicular horizon (DIXON 1994). However, a similar texture has not been found below the pavements of the study area. Instead the observed fine material is commonly very loose and powder-like and contains an extraordinary amount of $CaCO_3$ (Fig. 194).

Fig. 194: Accumulated subsurface dust of mainly carbonatic composition, likely originiating from eolian input.

Fig. 195: Dust from dirt road (National Road No. 52) traversing the study area.

In all cases observed, the parent material of desert pavements and associated carbonate accumulation horizons have been fanglomerates free of any carbonate rocks. The main lithological components are schists and phyllites. Therefore an autochthonous source of the $CaCO_3$ can be excluded. A fluvial transport of the carbonate material can be ruled out due to the catchment lithology of most observed pavments being free of any carbonate rocks. As a conclusion the carbonate must have accumulated as a result of eolian input.
Several sources of eolian input are likely. Remarkable amounts of dust get suspended from National Road No. 52 by the enormous number of 2,000 trucks per month passing through the study area (Fig. 195), but this dust should not have any significant carbonate content due to the carbonate-free road material. Flooding and fluvial activity on the floodplain has been observed to cause the deposition of remarkable amounts of loam-like material. This material should theoretically contain small amounts of carbonate rock as well, because outcrops of limestone and carbonate rock are known from several places of the study area.

However, these outcrops are estimated to make up for less than 3 % of the entire outcrop area of the study area, which reduces the possibility of significant carbonate input from local sources.

Dust storms of regional to supraregional extent are probably the most likely source of carbonate input (RUTHSATZ 1977). These storms have been reported to be active particularly during the winter season, delivering material from as far as Chile to the Cordillera Oriental. Dry, winterly westwinds are essentially caused by the subtropical pacific high pressure cell and are referred to as zonda (ENDLICHER 1995, WEISCHET 1996). The observed slight dust cover on some of the pavement clasts may point to repeated input by dust accumulation.

Even without definite evidence, the carbonate below the observed desert pavements is likely to have been added by eolian input. Therefore a close genetic link between the accumulation of carbonate by dust input and the formation of desert pavement is quite possible. While dust with a certain carbonate content gets trapped by the rough surface. This results in a slow upward displacement leading to the concentration of pebbles at the surface. This mechanism described as cumulic pedogenesis by MABBUTT (1979) does not entirely exclude surface deflation or wash processes but emphasizes the importance of dust input for the subsurface accumulation of fine material which in turn leads to the differentiation and segregation of grain sizes.

This interpretation implies that even though the areas covered with desert pavement are apparently inactive surfaces, slow pedogenic processes continue to accumulate carbonate and fine material from eolian input. Therefore the desert pavements can be interpreted as part of the present geomorphological system. The ongoing subsurface carbonatization is genetically linked to these desert pavements. As mentioned above (4.6.1.), it leads to the formation of a powder calcrete which could be regarded to represent the first stage of calcrete formation typical for semi-arid climates.

5. INTERPRETATION AND DISCUSSION

While the previous chapter mainly presented relevant results from geomorphological mapping as well as sedimentological and pedological analysis, this chapter aims to put these results into a context to each other and carefully reconstruct a landform and landscape history. On one hand, this means to set up a chronological order for the various events which changed and formed the landsape through time. On the other hand, this also provokes the question as to the causes of landscape change. Against this background, the results from the study area are discussed within the regional frame of existing landscape evolutionary data.

5.1. ESTABLISHMENT OF A LANDFORM HISTORY

The oldest remaining geomorphological features within the study area are clearly the paleoplanation surfaces on top of most mountain chains. These surfaces are evident from their flat topography. They are the last relics of a pre-Andean landsurface.

Tectonic forces of the various phases of uplift and deformation within the Andean orogeny became the dominant geomorphic control for landform development on a longer time scale. Particularly the regional drainage network still reflects in many ways the importance of tectonic and structural controls.

Erosional terraces give evidence of past phases of relative tectonic stability. They have been observed at considerable elevations as compared to modern topography and are therefore indicative for the intense erosional, but also deformational processes all landforms have been subject to in an active mountain belt like the Central Andes. Considering this, the relative lack of landforms from this stage of geomorphic evolution evidence is easily understood.

Following these older stages of landform history, a series of depositional terraces indicates a timespan of cut-and-fill sequences within the study area. In particular three alternating phases of erosion and aggradation corresponding to three generations of fluvial terraces (T-1 to T-3), followed each other within the study area. These cycles either reflect significant climatic shifts or they could have been the result of tectonic base-level changes.

Particularly the youngest cut-and-fill cylcle is very well documented in the study area. The corresponding remnants of terrace T-3 reach a thickness of more than 150 meters and stand for an intense phase of aggradation. While this phase manifests itself by aggradation in the valleys, it corresponds to equally intensive periglacial processes like frost cracking and frost creep as well as gelifluction and glatthang formation in the higher reaches of the study area.

Severe environmental changes must have taken place following this phase, which is reflected by various landforms. Lateral alluvial fans form on top of the terrace surface and meandering channel forms imply a fluvial system very different from today. However, a strong pulse of incision must have taken in, probably accompanied by intense badland formation within the less resistant sediments, and gullying reaching far up into the zone of former periglacial activity. Progressing incision and steepening gradients cause repeated landslides and most probably rock avalanches. The depth of incision is not apparent from todays landform assemblage, but it is very likely that maximum incision was followed by a new phase of floodplain aggradation, reaching a level roughly corresponding to the present floodplain elevation.

Several smaller phases of changing environmental conditions, particularly evident from the formation of colluvial slopes followed by incision and alluvial fan activity, have finally completed the appearance of the present landform assemblage.

Based on this very compressed history of landforms within the study area, mainly four essential phases of landscape evolution can be subdivided. All of these phases cover timescales significantly different from each other and reflect geomorphic processes of different magnitude. Due to the different quantity and quality of preserved geomorphological detail, interpretation of these phases of landform development regarding the will consequently put emphasis on different dominant geomorphic controls.

1. *Cycles of uplift and deformation* - appropriate timescale $10^6 – 10^7$ years
2. *Cycles of cut-and-fill sequences* - appropriate timescale $10^5 – 10^6$ years
3. *Upper Quaternary (last glacial cycle)* - appropriate timescale $10^4 – 10^5$ years
4. *Holocene to present modifications* - appropriate timescale $10^1 – 10^4$ years

5.2. ESTABLISHMENT AND DISCUSSION OF A LANDSCAPE EVOLUTION

Synthesizing the combined results from Chapter 4, and keeping in mind the different timescales and phases of geomorphic evolution, the landscape history of the study area can be summarized as follows.

5.2.1. EARLY UPLIFT AND DEFORMATION

The landscape evolution of the study area begins with the initial phase of the Andean orogeny which lifted the entire area above sea level and caused weak erosion of the flat-lying lowlands during Late Eocene to Lower Oligocene (SALFITY ET AL. 1996). As mentioned above, paleoplanation surfaces are the oldest relics of a pre-Andean land surface within the study area. In analogy to paleosurface remnants in Bolivia, they are assumed to be the result of erosional processes during mid-Late Miocene, at a time when the landscape was still dominated by low relief, not exceeding elevations of more than some hundreds of meters above sea level (KENNAN 2000). Fluvial and denudational erosion and subsequent aggradation formed the surfaces (GUBBELS ET AL. 1993), cutting rocks as young as Middle Miocene (KENNAN 2000). Therefore this process of surface planation is not equal to the formation of an etchplain by deep chemical weathering. It might rather be described as smoothing of a low relief landscape undergoing active uplift. These landscapes had already been subject to erosion and depositional processes in externally draining basins, mostly piggy-back basins, since around 16 –13 Ma BP (MARRETT ET AL. 1994). Despite several local, sometimes regional modifications during the Quaternary, the drainage network of the Cordillera Oriental developed parallel to the eastward-migrating uplift and thrusting (REYNOLDS ET AL. 1994). This way it was tied to important structural controls. Very likely, the Miocene to Pliocene drainage pattern already showed remarkable similarities to the present drainage network (KENNAN ET AL. 1997, BLOOM 1998). The reddish sandstones of the Chaco Formation correspond to fluvial deposition within a sedimentary basin integrated to the Miocene-Pliocene drainage network.

From Late Pliocene times onwards the dissection of the paleosurfaces commenced, possibly triggered by global climatic shifts (GUBBELS ET AL. 1993, KENNAN 2000), but certainly pronounced by increased rain shadow effects which concentrated precipitation along the eastern slopes of the Andes and led to the aridization of the highlands (KLEINERT AND STRECKER 2001). While the ongoing dissection, possibly retrocedent, further incised the drainage network, relief became increasingly pronounced and the isolated paleosurface remnants, which in most cases corresponded to thrust blocks, were sculpted within the landscape of the study area. Subsequent uplift and further thrusting were responsible for the varying elevations which the paleosurfaces show at present.

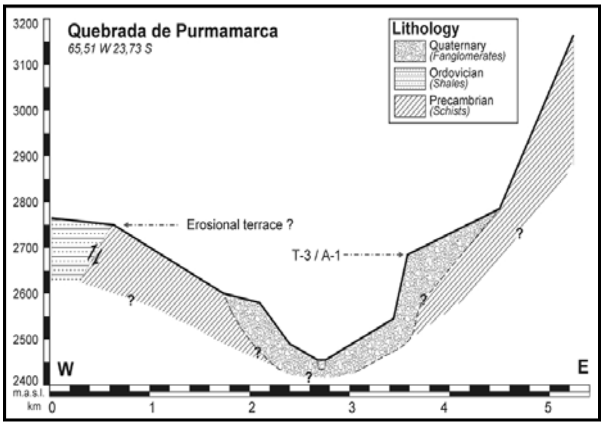
5.2.2. PLIOCENE-PLEISTOCENE CUT-AND-FILL SEQUENCES

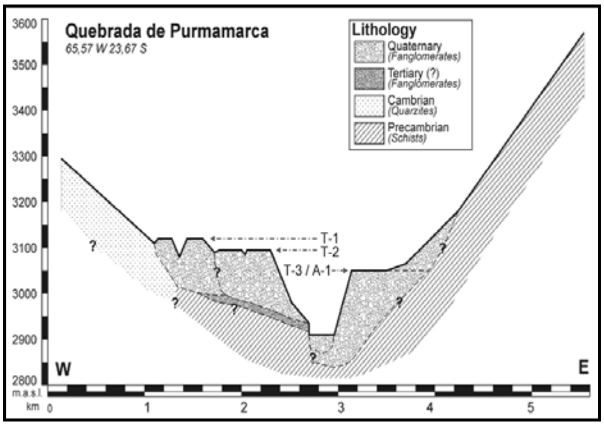
Within the study area the shift to semi-arid conditions, coupled with intensified uplift, caused a gradual but fundamental change of the sedimentary environment, coarse alluvial sedimentation replaced fluvial deposits. The study area had become subject to polygenetic cut-and-fill processes. Influenced by tectonic activity and/or climatic changes, several cycles of alternating erosion and aggradation have followed each other.

The earliest evidence for these processes is given by erosional terraces. Situated at relatively high elevations above the present valley floor, these isolated features account for the intense erosion and evacuation of sediment which has been going on at least since Pliocene times. Erosional terraces have been reported from several places in the Cordillera Oriental (WERNER 1984, TCHILINGUIRIANI AND PEREYRA 2001).

Sedimentary evidence from this time has been preserved at very few places of the study area as highly deformed and faulted fanglomerates and tilted lacustrine deposits, possibly corresponding to the Uquía formation, 2.78 to 1.5 Ma of age (MARSHALL ET AL. 1982). These alluvial sediments already indicate deposition in high-energy sedimentary environments under climatic and geomorphological conditions not very different from today.

Apparently, deposition of the Uquía formation was interrupted by another pulse of uplift and deformation during the Diaguita phase in Early to Mid-Pleistocene, followed by several tectonic events until at least younger than 1 Ma BP (SALFITY ET AL. 1984, HERNANDEZ ET AL. 1996, MARRETT AND STRECKER 2000). The intensity, regional extend and exact timing of these events is not known. Therefore a maximum age of less than 1 Ma BP can be assigned to all depositional terraces observed in study area. They do not show signs of severe tectonic disturbance and have essentially been preserved in their subhorizontal position. TCHILINGUIRIANI AND PEREYRA (2001) have come to a similar conclusion for the area between the Quebrada de Humahuaca and Salinas Grandes. Data from the Sierras Pampeanas (STRECKER ET AL. 1989) and the Bolivian Altiplano (WIRRMANN AND MOURGUIART 1995) point to relative tectonic stability in the Central Andes after 600 ka BP. For the Sierras Subandinas HERNANDEZ ET AL. (1996) even report an age of 250 ka BP for the last minor but geomorphologically significant tectonic activity.





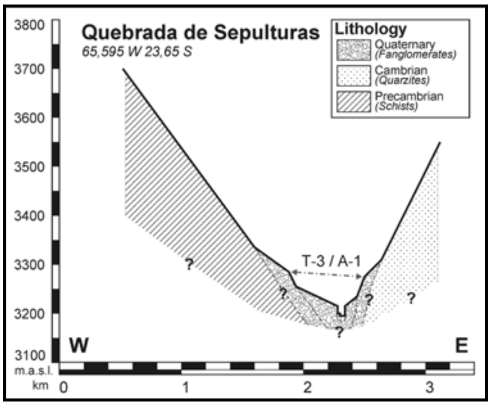


Fig. 196 – 198: Cross-sections through the Quebrada de Purmamarca at different locations

Based on topographic data, *three generations of fluvial terraces* (T-1, T-2 and T-3) have been distinguished within the study area (Fig. 196 - 198). Despite the chronological uncertainties mentioned above, they are thought to correspond roughly to the time interval of the last 600 ka. All of the terraces are built up essentially by coarse fanglomerates. This implies that alluvial sedimentation continued to be the major depositional process, most probably pointing to dominant semi-arid climatic conditions. As expected from their topographic location, the higher terraces T-1 and T-2 are markedly older than terrace T-3, which is particularly evident from their advanced stage of dissection and drainage channel evolution.

Two general models for the genetic history of the terraces can be inferred from the topographic association of the three terrace generations (Fig. 199). The first model suggests intense aggradation up to level T-1 and subsequent phases of alternating incision and stability, possibly coupled with lateral erosion. The second model however, proposes three alternating cycles of intense aggradation and incision. For several reasons this socond model very likely applies to the study area. At many places the deposits of terrace T-3 have been observed to laterally interfinger with deposits of the alluvial fan generation A-1. Activity of these fans initially commenced while T-3 deposition was still active. After T-3 deposition had stopped, A-1 alluvial fans were deposited onto the T-3 terrace surface in the entire study area, while they are absent on the terraces T-1 and T-2. This is evidence for an individual phase of erosion and subsequent valley fill for the terrace generation T-3. In addition, the terrace surfaces T-1, T-2 and T-3 are each inclined at different angles to different directions and different soils and lithologies characterize the terrace surface.

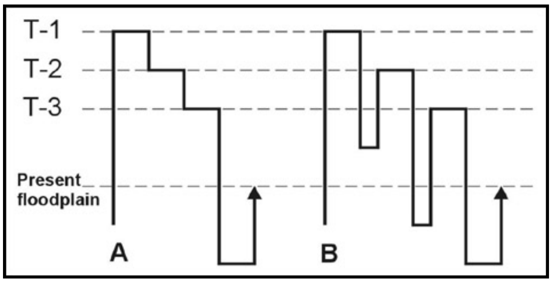


Fig. 199: Schematic sequences of terrace development inferred from topographic association; model B is preferred for the study area.

All of these observations supply evidence for an individual cycle of erosion, aggradation and subsequent incision of terrace T-3. Whether terraces T-1 and T-2 each correspond to a similar cylce of incision and aggradation ("cut-and-fill") cannot be decided on the base of available data. Considering the apparent age difference between the terraces and the manifold oscillations which have characterized the Pleistocene climate, three individual cut-and-fill events seem very likely.

In addition, evidence for a cyclicity of cut-and-fill events during the Quaternary has been reported from several locations within the Cordillera Oriental, whereas there is no agreement on the number of events. TCHILINGUIRIANI AND PEREYRA (2001) and WERNER (1984) distinguish one erosional and two depositional levels for the Quebrada de Humahuaca region, while AZAREVICH ET AL. (1999) report three depositional terraces from the Río Huasamayo at Tilcara. SCHWAB AND SCHÄFER (1976) and MARRETT AND STRECKER (2000) confirm five levels of Pleistocene depositional terraces younger than 1 Ma BP in the Quebrada del Toro. In any case, the fluvial system of the Cordillera Oriental seems to have been subject to enormous and intense oscillations during all of the Pleistocene.

Changes in fluvial behaviour of each stream usually go along with a change in at least one of the variables within the fluvial system (KNIGHTON 1998). Thus, there is a variety of causes for a stream to incise or to aggradate. T*ectonic activity* (e.g. uplift), *geomorphological changes* (e.g. base-level changes or sediment storage), *climatic-hydrologic change* (e.g.

increased/decreased discharge or sediment load, the so-called “Huntington principle”) and *human disturbances* are certainly the most common causes for incision (SCHUMM 1999). Considering the enormous intensity of cut-and-fill cycles and timescales involved, human disturbances can be ruled out as a major cause. Particularly climatic and tectonic causes have been discussed as relevant controls for the evolution of the fluvial system in the Cordillera Oriental.

Regarding the cut-and-fill cycles in the study area, the situation is conflicting. Significant tectonic activity and local uplift certainly have to be considered at timescales long enough to account for three cut-and-fill cycles. In fact, the two older terrace surfaces T-1 and T-2 have considerably higher inclination angles regarding their longitudinal profile compared to terrace T-3 (4,8-5° T-1/T-2 and 2,5-3,5° T-3). This might indicate a phase of minor uplift and/or thrusting in the upper study area postdating the deposition of T-1 and T-2 which could have resulted in a slight tilting of the terraces and subsequent erosion, setting the frame for renewed aggradation.
A tectonic control has been postulated for the cut-and-fill events of the five generations of depositional terraces in the Quebrada del Toro which are assumed to be the result of block uplift due to compressional forces (SCHWAB AND SCHÄFER 1976). Based on upstream divergence of fluvial terraces, COLOMBO ET AL. (2000) could show significant participation of fault activity in the evolution of an alluvial fan in the Precordillera of the San Juan province. Therefore minor tectonic events are likely to have occurred within the last 1 Ma years, even though the available data does not allow detailed conclusions regarding their effects on the evolution of the fluvial system within the study area.

The present semi-arid climatic conditions of the study area are mainly the result of the transitional position of the entire region between several climatic regimes. The dominant air masses reaching the area are of very humid tropical-continental origin, causing the present climate of winter rain (PROHASKA 1976, WEISCHET 1988). These circulation patterns are assumed to have been in place at least during all of the Pleistocene (GREENE 1995, HASELTON ET AL. 2002). While significant temperature changes are directly linked to global or at least hemispherical climate shifts during the Pleistocene and have therefore likely occured simultaneously beyond the limits of the Central Andes (ABRAHAM DE VAZQUEZ 2000), changes in moisture supply can be attributed to variations in the intensity and extension of the tropical climate system (GARLEFF ET AL. 1991).
The Pleistocene has been characterized by severe and frequent global climatic changes on the scale of several 100 ka (e.g. LOWE AND WALKER 1997). Multiple evidence for changing climates and environmental conditions throughout the Quaternary has also been reported from the Central Andes. CLAPPERTON (1993) divides at least four different phases of glaciation in the Bolivian Altiplano for the last 1.6 Ma. WIRRMANN AND MOURGUIART (1995) even report six different lacustrine phases for the same timespan in the Bolivian Altiplano. Nevertheless,

a regional extrapolation of a climatic interpretation of these lacustrine events to the Cordillera Oriental in NW-Argentina is problematic, as the existence of paleolakes in the southern Bolivian Altiplano does not necessarily depend on a climatic change effective in the entire Central Andes (FORNARI ET AL. 2001). For the time between 700 ka BP and 800 ka BP, coinciding with the Matuyama-Brunhes magnetostratigraphic boundary, NABEL ET AL. (2000) have reported a shift from wetter to increasingly drier climate in the Pampa of central Argentina. In addition, at approximately 800 ka BP a major shift in periodicities inherent to the orbital parameters responsible for solar insolation took place. This shift in periodicities from 41 ka cycles to 100 ka cycles has been accompanied by intensified global glaciation and cooling (LOWE AND WALKER 1997). Against this background, climatic shifts intense enough to directly affect the morphodynamic conditions are more than likely to have occurred within the Pleistocene and could very well be interpreted to have caused the repeated cylces of cut-and-fill within the study area.

5.2.3. UPPER QUATERNARY LANDSCAPE EVOLUTION

While paleoenvironmental data for South America covering the entire Quaternary are still very scarce and fragmentary, more reliable data exist at least for the last glacial cycle. Global climatic changes, as well-documented from deep-sea and ice core data in Greeland and Antarctica, have evidently had some effect on the climate in South America (COLTRINARI 1993).

Therefore not only the absence of any tectonic evidence within the terrace deposits point to climatic change responsible at least for the last aggradation and filling of the valleys and their subsequent erosion. In addition, multiple geomorphological evidence for envirnmental changes have been observed. Besides geomorphological evidence from within the study area discussed below, the evidence of an important aggradational phase is also visible throughout the entire Cordillera Oriental, e.g. Quebrada de Humahuaca, Río Iruya, Valles Calchaquies (Fig. 200 and 201).

Fig. 200: Incised valley fill in the Valles Calchaquies close to Cachi at 2,300 m.a.s.l.; summit in the backround is approximately 6,400 m.a.s.l. (170 km SSW of the study area).

Fig. 201: Incised valley fill in the upper Río Iruya at 3,400 m.a.s.l. (110 km NNE of the study area)

As no exact age dates presently exist, these events of supraregional aggradation can only tentatively be correlated to the last cut-and-fill cycle in the Quebrada de Purmamarca. Considering the structural style of uplift and deformation prevailing in the Cordillera Oriental, it is at least hard to imagine that all of these regionally not related examples of severe aggradation and incision should be a result of regional uplift. Therefore climatic causes have been inferred as being responsible for the last cut-and-fill cycle in the study area represented by the terrace T-3.

In the study area, an intense phase of erosion must have preceded the aggradation of terrace T-3 as the terrace deposits rest unconformably on solid bedrock at many places. The phase of terrace aggradation must have commenced at even below the present level of the floodplain, as the base of the terrace deposits is not always exposed. Therefore, the overall thickness of accumulated sediment certainly exceeds 160 meters for terrace T-3. Similar to the present situation, sedimentation predominantly resulted from debris-flow deposition.
In most cases the individual debris-flow events can be identified as they correspond to a distinct layer with marked upper and lower limits. No information is available concerning the frequency and recurrence intervals of these events. At no place in the study area erosional unconformities or other evidence for significant time lags in between the single debris-flow events have been observed. This implies either very reduced fluvial processes under relatively arid conditions or a high recurrence interval for the debris-flow events. However, modern debris-flow deposition rates and recurrence intervals are on the order of several tens to several tousands of years (COSTA 1984). Referring to these data, it is extremely difficult to estimate the duration of the entire terrace deposition. Assuming average thickness of 1,5 meters for each debris-flow event and a terrace thickness of 100 meters, this would imply a timespan of several thousand to more than 100 thousand years. Therefore the need for precise age data becomes obvious.
However, in contrast to the present debris-flows observed in the study area, the debris-flow deposits of terrace T-3 show a remarkable thickness of each individual debris-flow layer and enormous clast sizes of up to several meters in diameter. Even though the mechanism of past and present debris-flows must have been alike, they differ in intensity. Therefore the environmental conditions can be inferred to have been significantly different from today. A higher debris-flow volume and intensity should imply two things. Similar to today, a pronounced seasonality would reduce vegetation and account for high mobility of weathering debris and soil. In addition, it would cause intense and short-duration precipitational events. The combined effect is thought to have triggered debris-flow activity. However, the observed thickness and enormous clast size typical for the terrace T-3 deposits can only be attributed to a much more efficient mode of clast production, most likely by increased physical weathering due to frost action. Therefore temperatures must have been considerably lower than today. At least within a global frame, ice core data (e.g. from Vostok, Antarctica, e.g. COLTRINARI 1993, LOWE AND WALKER 1997) indicate a severe reduction

of global temperatures by up to 9°C colder than today for the time between approximately 116 ka BP and 13 ka BP. This alone seems to give a first hint to the significant participation of frost action in past geomorphological systems of the study area.
Sedimentological analysis of the terrace deposits confirms that the major catchment of source material for this enormous aggradation was very likely situated in the northern study area, possibly corresponding to higher topographic elevations. This would point to an increased importance of frost action as well. In addition, the sedimentological and lithological characteristics of the debris-flow deposits show relatively little vertical or horizontal variation within the terrace deposits. Over the entire timespan of terrace deposition the catchment must have been constant. It must have been large enough to comprise rocks of different lithologies.

Despite the homogeneity within the terrace deposits, debris-flow deposition has been interrupted by two phases of enhanced fluvial and lacustrine deposition. Evidence for these phases has been found at various places of the study area in the form of fluvial and even lacustrine sediments interfingering with debris-flow deposits. While the overall aggradational tendency continued, these phases are interpreted as phases of increased humidity. Particularly the second phase is of some importance, as it is evident in 10 of 15 sedimentological profiles within the Quebrada de Purmamarca. Within all profiles of the lower study area these fluvial and lacustrine deposits have been observed very close to the top of the T-3 terrace sediments indicating a severe change in depositional environments. Above these dominantly fluvial and lacustrine deposits, sedimentation shifted to minor debris-flow activity with lithological characteristics, depending on local geology. These deposits have been correlated with the oldest generation of alluvial fans in the study area (A-1). From interfingering terrace and alluvial fan sediments the transition from terrace deposition to local alluvial fan sedimentation has been inferred to have occurred very gradually.
In support of these information, a radiocarbon age within this phase could be determined to 49 550±1700 years BP. While this radiocarbon age can be considered reliable (Appendix), the calibrated calendar age would most probably be several thousand years older. Considering the applied temporal scale of investigation this age provides a good chronological date to go by. It indicates the cessation of intense debris-flow deposition in the entire Quebrada de Purmamarca and the gradual transition to local alluvial fan activity at around 50 ka BP.
As mentioned above, the enormous intensity of debris-flow deposition indicates significantly reduced temperatures within the study area. This is confirmed by a variety of geomorphological observations leading to the conclusion that large parts of the study area were subject to intense periglacial activity before or during the deposition of T-3. From active gelifluction and glatthang formation, the present lower limit of discontinuous permafrost is estimated to approximately 4,400 m.a.s.l..

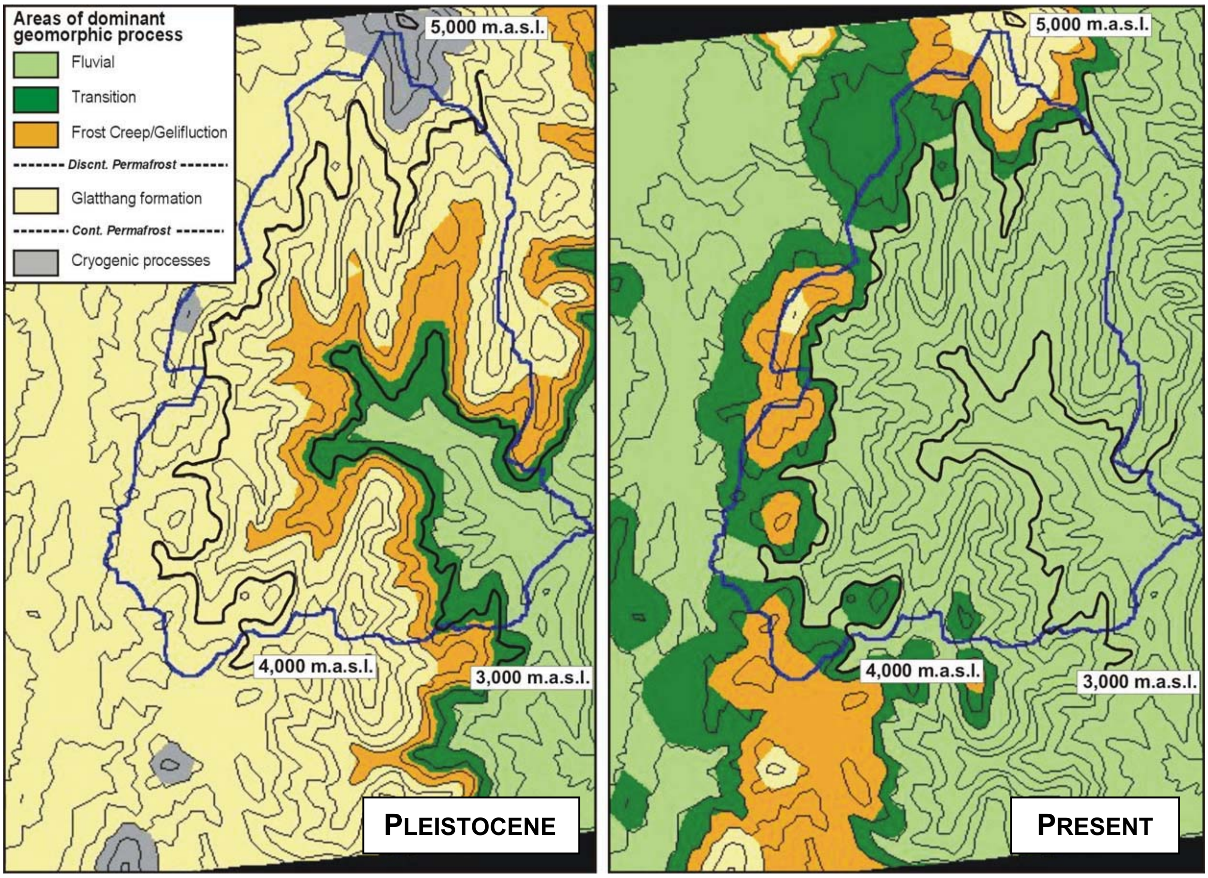


Fig. 202: Approximate shift of geomorphic zones Pleistocene – Present, induced from geomorphic evidence within the study area; note the expanded areas being subject to periglacial activity in the Pleistocene and consider the effects on debris production (contour lines highlighted).

A comparison between active and relict glatthang relief as well as a variety of other relict periglacial landforms down to elevations of 3,100 m.a.s.l. indicate a severe drop of the periglacial belt in the past by more than 1,200 meters, a fact which would have had most significant consequences for sediment production and transport capacity. Of the 414 km^2 large study area, 80 % would have been subject to frost creep and glatthang formation: a huge difference compared to only 7 % at present (Fig. 202). This implies an increased quantity of debris production by a factor of at least 10! In the study area this drop of the periglacial belt must have taken place before or during the deposition of terrace T-3. In contrast to the higher and older terraces, terrace T-3 does no show any signs of periglacial reshaping. Therefore periglacial activity of considerable intensity and duration has to have prevailed in the study area before about 50 ka BP, resulting in the enormous aggradation of the fluvial system. Despite the absence of modern analogues, the precise mechanisms of debris-flow generation has been assumed to be the combination of two preconditions. First of all the intense periglacial production of frost debris contributes material of all grain sizes (clay ? to boulders) from the slopes directly onto the floodplain. Intense precipitation and / or snow melt generates catastrophic flooding, probably intensified by very low infiltration rates on the frozen ground. These powerful floods picked up the abundant sediment, thereby transformed to debris-flows and eventually redeposited the material on the floodplain.

Observations from Humuhuaca (3,300 m.a.s.l.) by EBERLE (2000) seem to confirm an intense phase of accumulation. Although interrupted by at least on phase of soil formation, he mentions a phase of significant accumulation which started prior to 100 ka BP and ended around 35 ka BP. Interruptions by soil formation might very well be correlated to lacustrine and fluvial phases apparent from the sedimentological data of the terrace deposits. The postulation of severe shifts of the periglacial belt is supported by GARLEFF AND STINGL (1985), ZIPPRICH (1998) and ABRAHAM DE VAZQUEZ ET AL. (2000) who report a significant drop of mean annual temperatures of 6 – 9°C in the Andes of NW-Argentina for the last glacial maximum. For the pleniglacial timespan before 50 ka BP ZIPPRICH (1998) reports at least two phases of glaciation in the Sierra de Santa Victoria, a fact which implies generally low temperatures during the entire glacial cycle as compared to today. In addition, CARIGNANO (1999) reports substantial accumulation of loessic silts under arid and cold climate in the Argentine plains of the Córdoba province prior to 50 ka BP.
All of these observations confirm the general assumption of a significantly colder and drier climate in the study area before 50 ka BP and support the idea that the enormous depositional terraces are predominantly the result of periglacial debris production. However, an alternative hypothesis has been proposed by SEGEMAR-ITGE (1998). Without going into any detail, they attribute the increased debris-flow frequency to a formerly increased mean altitude of cloud condensation.

While implying a progressing reduction in sediment supply, the gradual transition from terrace deposition to A-1 alluvial fan activity might be attributed to slightly increasing temperatures. As mentioned above, the terrace deposition can be assumed to have taken place under significantly colder periglacial conditions. A slight increase in mean annual temperatures would have two effects. First of all, the area of the periglacial belt and amount of debris produced by intense periglacial activity in the upper study area would have bene reduced, leading to an overall reduction of sediment supply and enhanced fluvial processes. This is particularly true for the tributary quebradas. In comparison to the main quebrada these quebradas are located at higher elevations. Therefore an upward shift of the periglacial belt would gradually have accentuated fluvial activity in each of these tributary quebradas. Nevertheless, a huge amount of sediment is available as periglacial hillslope debris which covers the extended slopes of former periglacial belts. While a high climatic seasonality with intense precipitational events continues to be characteristic, the tributary quebradas gradually developed alluvial fans. As expected this phenomenon commenced later and lasted longer in the upper and higher parts of the study area, because here periglacial conditions prevailed the longest and debris production continued to be active.

Following the deposition of these A-1 alluvial fans a severe and pronounced phase of incision commenced in the study area. The most obvious evidence for this phase are the almost vertical walls of terrace T-3 and alluvial fan A-1 deposits of sometimes more than

100 meters in height. The incision of the floodplain and the removal of enormous amounts of sediment must have occurred at relatively high rates because of the enormous height and steepness of the vertical terrace walls. It continued to a level even below the present floodplain. This is indicated by several landslide deposits which are presently being buried or mantled by present floodplain deposits. These landslides correlate very well with this enormous phase of downcutting. The incision operated at rates which did not allow for slope adjustment. This oversteepening led to a general instability of the valley walls, both in solid rock and terrace deposits which made the slopes prone to failure and mass wasting.

Several considerations have to be taken into account when it comes to determining a possible reason for this incision. The evacuation of the exraordinary coarse sediments requires considerable transport capacities. While debris-flows are known to have remarkable transport capacities, they usually do not erode material from the floodplain. Therefore the processes responsible for incision are likely to have been flows of comparetively low sediment concentration and high water volume and velocity. These are the high-magnitude and low-frequency events discussed by WOLMAN AND MILLER (1960), pointing to a high seasonality of the precipitation events. A pronounced dry season is very likely, because a well-balanced and constant humidity would have created a vegetation cover much too dense for intense linear incision. This problem had long been recognized by LANGBEIN AND SCHUMM (1958) who argued that the effect of a climate change does not only depend on the direction of the change (e.g. from dry to wet), but is particularly controlled by the climatic conditios before the change. Therefore two factors favor pronounced fluvial processes as needed for incision.

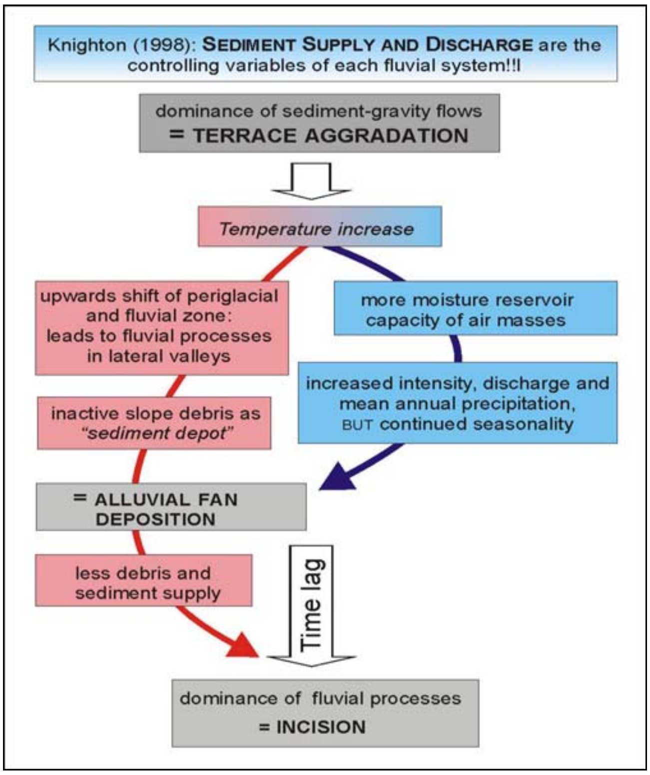


Fig. 203: Schematic model of climatically controlled terrace aggradation and incision.

First of all, an increase in mean annual temperature, even if only by a few degrees, would consequently lead to an upwards shift of the periglacial belt and to a general reduction of sediment production. It would pronounce fluvial and alluvial activity at the increasingly higher elevations including the tributary quebradas formerly choked by extraordinary sediment production. Nevertheless, a significant amount of debris remains readily available on the slope areas. This led to the increasing importance of the A-1 fans in the study area. Secondly, an increase in temperature is more than likely to have had important effects on the overall intensity of the tropical circulation as well as on the the regional and local capacity of the air masses as moisture reservoirs. As evident from basic meteorological rules, warmer air masses have the capacity to take up significantly higher amounts of moisture and release this moisture in precipitational events of increased intensity. While precipitational events continue to be highly seasonal, this would cause a further pronounciation of the intensity of each event and lead to overall higher mean annual precipitation rates.

The combination of reduced sediment supply and enhanced precipitation is thought to have produced flooding events powerful enough to erode, transport and remove the enormous amounts of terrace deposits out of the study area. Once the volume of sediment supply falls short of a critical threshold, an almost impulsive incision sets in. Assuming high bedrock incision rates of up to 15 mm/yr (PRATT ET AL. 2002), an incision of 100 meters would only take about 6,670 years. However, assuming similar discharge, the low resistivity of the unsonsolidated terrace deposits compared to solid bedrock might result in even higher rates.

When applying this model to the study area, two further considerations have to be mentioned. The considerable amounts of frost debris resulting from the lowered periglacial belt are thought to have produced thick sheets of debris covering the slopes in the study area. These enormous masses of debris are likely to have provided sediment to the fluvial system for a relatively long timespan after the increase of temperature. Therefore the onset of incision must have occurred only with a certain time lag during which alluvial fans of generation A-1continued to be active. A very similar mechanism of impulsive alluviation has been recently proposed by PRATT ET AL. (2002). They postulate a climate change towards wetter, monsoonal conditions as the trigger for changing controls of the geomorphological and fluvial system, but report the onset of incision with a time lag of a few thousand years. During this time, preexisting hillslope material was liberated, leading to intense sedimentation. Incision only commences when the bulk material of available hillslope debris has been exhausted.

Even though no detailed field data exist, the marked difference of hillslope debris thickness between the upper study area and the lower study area coupled with the areally intense and advanced dissection of most hillslopes might serve as preliminary evidence for the above suggested model (Fig. 203).

The observed time lag between terrace deposition and incision might also point to different times of temperature and moisture increase. In this case, an increased mean annual precipitation going along increased storm intensity and higher stream power would lag behind a temperature increase leading to the shift of the periglacial belt. Particularly with regard to the glacial advances in the Central Andes, some authors have reported a significant time lag of about 4,000 years between the maximum temperature depression and the maximum snowline depression (e.g. ZIPPRICH 1998). Again, this indicates that increased moisture supply is unlikely to have occurred during maximum temperature depressions. This phenomenon can certainly be attributed to an increased reservoir capacity of warmer air masses, but is most probably explained by intensity changes coupled within the tropical if not global circulation system. Therefore, the question why despite significantly higher temperatures no *present* incision has been observed might be explained by an increased overall aridity within the Central Andes due to processes tied to the global atmospheric circulation.

The considered increase in temperature necessary to trigger a change in the geomorphological system has to be seen within the appropriate frame. As mentioned above, before about 50 ka BP the temperature can be assumed to have been significantly lower than the mean annual temperatures today, possibly on the order of 6° to 9°C. For the following phase of alluvial fan deposition and subsequent incision, temperatures did certainly not reach today's values. Nevertheless, temperatures 3° to 4°C colder than today have been assumed for the period between 40 ka BP and 25 ka BP for the Santa Maria Basin, about 300 km south of the study area (BOOKHAGEN ET AL. 2001). This would have corresponded to an increase of 10-15 % for the mean annual average precipitation.
In any case the observed intense incision can be attributed to enhanced fluvial conditions, most probably due to an increase in mean annual precipitation, even though still coupled with a strong seasonality and a pronounced dry season typical for most tropical monsoonal climates of winter rain. Even if the model described above gives an explanation for the mechanism of incision it does not reveal the timing of events.

As mentioned above particularly the steep terrace walls, multiple landslide deposits but also intense slope dissection postdating A-1 alluvial fans reflect the postulated fast and deep incision of the Río Purmamarca within the study area. Intense downcutting has been reported to be responsible for numerous landslides and associated landslide-dammed lakes in all of NW-Argentina (TRAUTH AND STRECKER 1999, STRECKER AND MARRET 1999, HERMANNS AND STRECKER 1999, TRAUTH ET AL. 2000). Although most of these authors consider earthquakes as possible triggers for the actual landslide event, they emphasize that an oversteepening of the slope is a necessary precondition for landsliding. This oversteepening is attributed to downcutting and lateral undercutting by fluvial processes during a phase of wet climatic conditions between 40 ka BP and 25 ka BP. Further evidence for increased moisture supply is given by ZIPPRICH (1998) who mentions a phase of glaciation from the Sierra de Santa Victoria around 30 ka BP. CARIGNANO (1999) reports the onset of intensive soil formation in the Argentine plains for around 50 ka BP. In addition, a well established phase of high lake levels in the southern Bolivian Altiplano is assigned to an intensified tropical circulation and moisture supply from the east prior to about 21 ka BP (WIRRMANN AND MOURGUIART 1995). This phase called the *Minchin lacustrine event* is likely to have commenced around 40 ka BP but evidence for repeated moist intervals extends back to 72 ka BP (FORNARI ET AL. 2001).

Within the study area various evidence has been observed to postdate the accumulation of the terrace deposits. A reddish, clay-rich paleosol interpreted as a cambisol has been found on top of most terraces underlying all other paleosols and soil horizons. While the clay enrichment implies higher amount of overall moisture and enhanced plant growth, the intense reddish color reflects a persistent dry season.
At many places incised meanders are found on top of the terrace surface of T-3. This might reflect the fact that the initial fluvial system, which formed as a direct result of the upwards shift of the periglacial belt and stopped terrace deposition, had to cope with enormous quantities of sediment load. The thick sheets of hillslope debris served as a source for grain sizes larger than silt. Due to the high sediment supply the stream's energy (stream power) was used predominantly for transport and distribution of the sediment. The channels formed meanders and only when sediment supply decreased, incision of the meanders commenced. Alternatively, the tendency of a river to meander usually indicates a low river gradient, a relatively stable floodplain due to an increased percentage of fine grain sizes and lower flow velocities (KNIGHTON 1998). This would much rather point to wet conditions, e.g. by intensified chemical weathering, or a denser vegetational cover. Whether this can be considered the result of a slight increase in mean annual temperature and precipitation or corresponds to a significantly reduced seasonality is a crucial question for the paleoclimatic interpreation of these forms. Nevertheless it is difficult to answer, as long as there is no general agreement on the causes for river meandering (KNIGHTON 1998).
Further evidence for increased moisture might be indicated by tafoni and abri forms in andesitic rocks. Even though destructional processes are currently active on the interior walls of the abri enlarging the abri forms, these processes are not thought to have initialized the formation of tafoni and abris. Their original formation is attributed to an enhanced moisture supply. Whether they correspond to the same phase of moister environmental conditions as the soils and meanders mentioned above cannot be determined due to their isolated geomorphological location.

Summing up, all of these observations corroborate the existence of an important phase of increased humidity within the study area. As mentioned above, several authors have confirmed the existence of the *Minchin lacustrine event* for the Cordillera Oriental and the southern Altiplano in the Central Andes. Even though it is evident that this wet phase postdates the terrace deposition as well as the alluvial fan generation A-1, the question remains whether the intense pulse of incision was chronologically linked to this wet phase, or if the intense incision observed in the study area occurred much later.

In order to answer this question, theoretical considerations lead to a models shown in Figure 204. Provided a parallel increase of temperature and precipitation the transition into a wet phase would gradually result in a much denser vegetation, which would significantly decrease erosional capacity. Assuming an increase in precipitation preceding that of temperature, it would be impossible to explain the obvious shift from terrace accumulation to alluvian fan depostion by an upwards shift of the periglacial zones. But if a temperature increase preceded an increase in precipitation, precipitation would remain constant, while the periglacial belt moved upwards. Coupled with an increasingly exhausted sediment supply, an initial increase in precipitation would have triggered short but effictive incision, before the complete shift to wetter conditions typical for the Minchin period would have ended incision due to denser vegetation (Fig. 204). Only this model, where incision seems to be a phenomenon characteristic for a transition between two contrasting climatic phases, combines all observed landforms and assigns them a relative chronological order. In addition, a significant time lag between temperature and moisture peaks has been reported from various authors (e.g. ZIPPRICH ET AL. 1999) and confirms this interpretation. As indicated above, this time lag might be attributed to relatively slow shifts within the global atmospheric circulation.

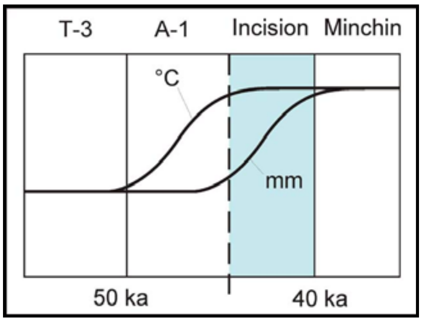


Fig. 204: Schematic model (no scale) explaining the sequential landscape development from T-3 terrace accumulation over alluvial fan activity to incision and the subsequent wet phase.

Keeping in mind the model described above, the incision would have preceded a substantially wet period like the Minchin lacustrine phase, characterized by relatively stable conditions allowing soil formation, stable stream banks with a tendency to meander and the development of tafoni and abris. Even though all of this evidence points to enhanced chemical weathering, possibly coupled with an increased density of vegetational cover, the climatic conditions should not be referred to as entirely humid. As indicated by reddish soils resulting from rubefication processes, a seasonality with an inherent dry phase was still present. This dry season might have been reduced going along with an increased overall annual precipitation. While the existence of the Minchin lacustrine phase is in fact supported by the field evidence, its relation to incision is still not certain.

At many places in the study area the reddish soils are directly overlain by a cemented sand horizon. Macromorphological and micromorphological analysis support the idea that these sand crusts can partly be attributed to eolian activity and fluvial reworking. The existence of a relict sand ramp could not be prooven, but seemed likely on the base of available data. Therefore it can be argued that significant aridity must have characterized the entire region. Considering the absence of eolian forms, semi-arid to arid conditions in the study area, aridity must have been even more intense, possibly in combination with significantly reduced vegetation cover, accounting for the high mobility of the fine and medium sand.
A significant drop of temperatures by 5° to 9°C (GARLEFF AND STINGL 1985) and the absence of glaciers in the Sierra de Santa Victoria (ZIPPRICH 1989) seem to give evidence for dry and cold conditions between 22 ka BP and 18 ka BP, typical for the Late Glacial Maximum (LGM). This drop of temperatures has been expected to cause a drop of the periglacial belt similar to the phase before 50 ka BP. However, no evidence from the study area for periglacial activity postdating terrace deposition has been found. Neither the terrace surfaces nor the terrace slopes show any sign of periglacial processes or reshaping. Nevertheless, a reactivation of relict hillslope debris or even renewed periglacial acticity and debris production cannot be excluded. If there had been any periglacial production of slope debris, a renewed phase of subsequent dissection must have removed any evidence.

The deposition of the sand crust was evidently favoured in preexisting depressions like drainage channels and below rims. Therefore it is assumed to postdate a first phase of incision. Interestingly, the sand crust has not been observed below the level of the T-3 terrace surface anywhere in the study area. Instead, at many places the crust has obviously been followed by processes of dissection and erosion. This crucial fact clearly provokes two considerations.
Either the intense first phase of incision postdated the deposition of the eolian sand crust and occurred later than assumed above, or two phases of relatively strong incision have taken place, with one of them preceeding the LGM and the other one postdating it. The existence of the sand crust within a soil profile associated to the A-1 alluvial fans would certainly support the thesis of an important phase of incision after the LGM and implies that the major dissection and erosion of the A-1 alluvial fans, making them entirely inactive landforms, is indeed younger than the sand crust of approximately 20 ka BP.
In analogy to the above presented model, again it could have been a transition to wetter conditions triggering severe incision. However, this implies two options. First of all, the above-mentioned Minchin lacustrine phase could have been without any effects on the fluvial system in the study area. The only evidence for this wet phase within the study area are the reddish soils underlying the sand crust. In addition, the meanders imply wetter climate before the deposition of the sand crust. Again, this would impose the question for the mechanisms of incision, as it would contradict the above-mentioned model. It would imply a relatively short climatic shift, possibly too short for severe incision to occur.

Alternatively the overall climatic conditions might have been too wet for any incision to occur as a vegetation cover much denser than today had developed. This would mean that the Minchin wet phase differed significantly from any climatic conditions since then. Possibly, evaporation rates were markedly lowered due to the overall colder climate, allowing intense plant growth and restricting linear incision. This assumption is tentatively confirmed by the absence of a second reddish soil. In this scenario, conditions accounting for increased stream power and incision would have prevailed only during a transition from arid to semi-arid, or possibly subhumid conditions, but not during transition to a very humid phase like the Minchin .

Provided that two phases of incision had removed and evacuated the large amounts of sediment from the study area, the first of them occurred before, and the second one after the LGM. Both of them could be explained by the above-described model, where incision occurs at the transition to wetter climate. The only prerequisite would be the existence of two markedly wetter phases in the study area. The lack of a second reddish soil could be explained either by a much shorter duration of the second wet phase or by a different intensity of moisture supply.

Several generations of calcretes have been observed on top of most A-1 alluvial fans, overlying the sand crust. The formation of such carbonate crusts hase been attributed to a range of 150 – 500 mm of precipitation per year (EITEL 1999). Considering this remarkable range, semi-arid climatic conditions might indeed have experienced a certain increase of mean annual precipitation, but this tendency of increased moisture did not reach the level of the Minchin lacustrine phase. Therefore no reddish soils could be formed due to a prevailing seasonality. Only the overall mean annual precipitation might have been increased, leading to a accentuation of discharge events.

In fact, several authors report a late glacial phase of relatively increased humidity from the Central Andes. ZIPPRICH (1998) has identified a glacial advance around 16 ka BP in the Sierra de Santa Victoria. SYLVESTRE ET AL. (1999) report high lake levels in the southern Bolivian Altiplano between 16 ka BP and 12 ka BP, a phase called *Tauca lacustrine phase*. Lake levels, however, do not necessarily reflect the regional climatic situation (FORNARI ET AL. 2001) and therefore wetter climatic conditions in the Cordillera Oriental of NW-Argentina are more likely to have occurred around 16 ka BP. ZINCK AND SAYAGO (2001) report multiple cycles of soil formation from the Andean northwest, clustering between 20 and 14 ka BP. As a consequence, a second phase of incision could very well be assumed for the time between the LGM and 16 ka BP which has finally dissected the A-1 alluvial fans and has at many places cut through the slopes and the sand crust. Incision must have been intense, probably reaching down to below the present floodplain.

Summing up, it cannot be decided when the major part of the incision and evacuation of the sediment out of the study area has occurred, but it seems likely that particularly the transition to the second lacustrine phase, which is thought to have taken place during late glacial times throughout the Central Andes, eventually resulted in severe incision.

Due to this uncertainty, only a minimum age can be assigned to the initial formation of the A-2 alluvial fans. These fans have adjusted approximately to the present floodplain and are therefore interpreted to postdate the last phase of intense incision. In contrast to the A-1 fans they hardly show any dissection. In addition, they do not exhibit any evidence for periglacial activity. Due to their present activity this does not necessarily reveal any information about their age. In any case the onset of the A-2 alluvial fans marks the end of intense incision and aggradation. Depending on the timing of the incision, it is likely that these fans have started deposition after about 16 ka BP under climatic conditions relatively similar to today. Their deposition is thought to have followed or accompanied a phase of floodplain aggradation, which must have taken place after the intense incision. While most landslide deposits observed in the study area have been interpreted as the result of a phase of increased relief and slope instability due to incision and possibly wetter climate, these same landslides are presently being buried by floodplain deposits. Therefore several meters to tens of meters of aggradation must have occurred since then. In analogy to the present climatic conditions, floodplain aggradation might correspond to an episode of increasingly semi-arid to arid conditions, where the sparse vegetation cover makes more sediment available to flooding and debris-flow activity.

Even though no absolute age dates exist, the formation of the multiphase calcretes on top of the A-1 alluvial fans is assumed to have started at the same time. While activity on the A-2 fans was too dynamic to allow any soil formation, the preceding incision had left most A-1 fans inactive and prone to the accumulation of carbonates. At most places, these calcretes have experienced several phases of cementation and fragmentation since their initial formation. This implies, that late glacial to present climate has experienced several osciallations and has not been stable for most of the time, while it kept returning to semi-arid conditions similar to today. A polyphase colluvial cone in the Quebrada de Sepulturas, which has formed within a rock slide scar, confirms this assumption. The rock slide itself probably dates back to the phase of incision or to the initial phase of A-2 alluvial fans. The following formation of the colluvial cones can only have occurred under relatively dry conditions, but has been interrupted by erosion at least once.

On a global scale, high-frequency climatic changes were very common during late glacial times, and even in NW-Argentina a considerable cyclicity of climatic changes has been confirmed by paleosol sequences (ZINCK AND SAYAGO 2001). From the study area no pronounced cyclicity has been evident with the exception of the polyphase calcretes. However, the onset of A-2 alluvial fan sedimentation is characteristic for a semi-arid to arid transition from late glacial climate into the beginning Holocene. Very dry conditions known as the Ticaña event are also reported from the Bolivian Altiplano following the pronounced Tauca wet phase after about 12 ka BP (SYLVESTRE ET AL. 1999). ZIPPRICH (1998) and ZIPPRICH ET AL. (1999) report significant periglacial from the Sierra de Santa Victoria and infer dry and cold climatic conditions, probably similar to the central european Younger Dryas.

5.2.4. HOLOCENE LANDSCAPE EVOLUTION AND CURRENT MODIFICATIONS

After floodplain aggradation and alluvial fan depostion had elevated the floodplain approximately to its present level, a renewed change of conditions must have taken place. This change is reflected by colluvial slopes at the foot of the terrace walls. Along the floodplain, hese slopes show a vertical scarp several meters high. Presently they are subject to dissection by A-3 alluvial fans. The formation of these slopes has been attributed to dominant slope wash processes under relatively humid and stable conditions, restricting strong incision and erosion. In fact most paleoclimatic data for NW-Argentina and Bolivia indicate increased moisture supply between 10 ka BP and 8.5 ka BP (MARKGRAF 1984, FERNÁNDEZ ET AL. 1991, J.J. KULEMEYER 1998, SYLVESTRE ET AL. 1999, ZIPPRICH ET AL. 1999).
However, the colluvial slopes are likely to have adjusted to the former floodplain. Therefore the existence of a vertical scarp in their distal part seems to be evidence of enhanced subsequent lateral erosion. While implying some importance of fluvial processes, lateral erosion can neither be explained by humid conditions accounting for a certain stability, nor by semi-arid conditions comparable to today. Therefore it is suggested to assign this phase of lateral erosion to a phase of intermediate humidity. Aside from the formation of colluvial slopes, this subhumid phase might also be responsible for the marked lateral erosion visible in the distal parts of the A-2 alluvial fans. Chronologically it present the direct climatic transition from humid climate between 10 ka BP and 8.5 ka BP to a renewed phase of seim-arid conditions. Alternatively it might correspond to Mid-Holocene times, as observations in different parts of NW-Argentina by MARKGRAF (1984), J. J. KULEMEYER (1998) and ZIPPRICH ET AL. (1999) point to climatic conditions slightly wetter than today between about 3,7 ka BP to 1,9 ka BP. This second scenario would imply a period of relatively stable geomorphological conditions without any evidence having followed the humid climate of the early Holocene. In fact, several authors have mentioned a particularly dry phase between approximately 8 ka BP and 4 ka BP (e.g. KULEMEYER 1998, ZIPPRICH ET AL. 1999). This remarkably dry phase, also known as the "archaeological silence" (no archaeological evidence for human occupation, J. A. KULEMEYER 1998), graded into more humid conditions between 4 ka BP and 3 ka BP, giving way to a more constant human occupation of the Cordillera Orienta of NW-Argentina (FERÁNDEZ ET AL. 1991, KULEMEYER ET AL. 1999).
Since then, the floodplain must have remained at a relatively constant level. At some point, the deposition of the alluvial fan generation A-3 commences, which continues to be active until today. Alluvial fan activity is intimately linked to the badland formation and gullying processes typical for the present semi-arid conditions, all presently observed to be active in the Quebrada de Purmamarca. The clear dominance of debris-flow versus fluvial activity has been demonstrated for several quebradas in the study area, leading to an overall tendency of floodplain aggradation. This tendency is thought to be a relatively young phenomenon. Possibly it has resulted from a constant increase of anthropogenic influence leading to intense gullying of the slopes and aggradation of the floodplains as proposed by J. J.

KULEMEYER (1998) for the Sierra de Santa Victoria. More likely, aggradation is the consequence of intensified or more frequent precipitational events leading to increased sediment supply. A relatively high variability is inherent to most semi-arid climates. Over the last several hundred years several rapid and intense climatic oscillations have been recorded by lake sediments in NW-Argentina (GROSJEAN ET AL. 1998). Even on time scales no longer than tens of years the quasi-periodic cycles of ENSO have been reported to be responsible for phases of high flooding frequency in the Arroyo del Medio, a tributary of the Quebrada de Humahuaca (MAAS ET AL. 1999). Therefore the study area can be considered to be a highly dynamic region, not only based on the observation of strong, presently active processes, but also taking into account the variability of climatic conditions which can easily change the morphodynamic situation, not only within the study area.

Reacting to the above-mentioned aggradational processes inherent to the present morphodynamic state, people have started to directly influence the landscape evolution of the study area in order to prevent damage by debris-flows or floods. At several places fences, levees and artifical drainage channels give of this effort and can be regarded as the ultimate element of the landscape evolution in the Quebrada de Purmamarca.

Considering that these highly dynamic conditions are typical not only for the Quebrada de Purmamarca, but for the entire region of the Quebrada de Humahuaca, short-term solutions are probably not the appropriate way to prevent future damage. While some authors ironically propose the upslope transfer of complete settlements (AGUERO 1986), an increased effort to understand the complex geomorphological system and subsequent recognition of areas of particularly high risk potential (e.g. SOLIS AND OROSCO 1996) would certainly reduce much future damage.

6. CONCLUSION

This study has aimed to decipher the landscape evolution of the Quebrada de Purmamarca by combining geomorphological, sedimentological and pedological data. This integrative approach has not only multiplied the amount of available information and detail, it has also served to show the interconnections between the various geoscientific disciplines. Although landscape has traditionally been characterized as an assemblage of individual landforms, this study has emphasized the character of landscape as an open system. Only an integrative approach can therefore meet the wish to understand controls and processes coupled within this system.

Based on the available geomorphological, sedimentological and pedological data, the study has arrived at the division of four essential episodes for the landscape evolution of the Quebrada de Purmamarca. Although at different temporal and spatial resolutions, geomorphological and sedimentological evidence has been revealed for all of them. The evolution of the landscape commences with the initial uplift of the Andes and continues in a series of intense modifications by erosional and depositional sequences covering millions to hundred thousands of years. Over longer timescales like these, tectonic activity accounts for major modifications of the landscape. Consequently, landscape evolution and geomorphology are never free of an important geological aspect.
However, on shorter time scales of up to a few hundred thousand years, geomorphological and sedimentological analysis has uncovered an extraordinary variety of data in the study area. Many of the observed results could be interpreted as the result of significant environmental changes in the past, which had severe influences on the entire geomorphic system. In particular the deposition of the large terrace systems characteristic for many valleys within the entire Cordillera Oriental has been shown to result from very cold and probably arid conditions during a long times pan within the last glacial cycle. Very likely the remarkable intensity of incision can be attributed to environmental changes towards climatic conditions less cold and probably less arid than today. In addition evidence for episodes of climates considerably wetter than today has been detected. Therefore in most cases intense *climatic changes* have been suggested as being the dominant controls of Late Quaternary landscape evolution. Against this background it becomes obvious that particularly the integrative geomorphological approach is capable of producing a much more detailed picture of past environmental changes than some of the traditional, sometimes very descriptive geomorphological concepts have done.
Even though the remarkable number of observed landforms and soils as well as sedimentological data is certainly indicative of frequent climate changes, the establishment of a relative chronological order of events is not always unequivocal. In many ways the results from the study area correspond to global paleoclimatic trends, inferred for example from ice core data. Even within this dynamic high-mountain environment manifold

variations within the geomorphic system at relatively short time scales have been detected. The existence of important periglacial activity certainly goes along with the global trend of astronomically induced temperature reduction during most of the last glacial cycle. In addition the detection of several significant environmental changes within an overall glacial period is consistent with the large number of interstadial events characteristic for the Late Glacial times. Landscape does record most of these oscillations and geomorphology has the capacity to detect and interpret them.
Particularly with regard to global as well as regional paleoclimatic interpretation, however, the establishment of an chronological frame is undispensable to allow extrapolation, comparison to related research and integration of the results into a global frame. In this respect this study has contributed a preliminary result. The phase of enormous aggradation of the thick terrace systems has been shown to be significantly older than the previously assumed Late Glacial Maximum age. However, concerning the timing of their subsequent incision, no definite decision has been possible as yet, based on the available data. Therefore landscape evolutionary studies and considerations would certainly benefit from future advances in relative and absolute age dating methods. Particularly with regard to radiometric dating techniques and tephrochronology, but also concerning palynological and related microbiological methods, the Quebrada de Purmamarca and its sedimentary landforms offer manifold possibilities for further research.

Having shown the enormous past and present dynamics inherent to the semi-arid high-mountain catchment of the Quebrada de Purmamarca, the study reveals urgent implications regarding regional development. To a certain extent, the present morphodynamic situation of the study area as well as large parts of the Quebrada de Humahuaca has already led to an awareness for potential environmental risks among the people of the region. Due to the inherent large inter-annual variability of the geomorphic processes typical for the region, much more catastrophic events are likely to occur within the immediate future and have evidentially occurred over long times in the past. Therefore this study also helps to illustrate the highly dynamic character of the Quebrada de Purmamarca. This way it can hopefully emphasize the need for further research concerned with both the past and the present geomorphological processes.

ABSTRACT

THE QUEBRADA DE PURMAMARCA, JUJUY, NW-ARGENTINA: LANDSCAPE EVOLUTION AND MORPHODYNAMICS IN THE SEMI-ARID ANDES.

The aim of the presented study has been the reconstruction of a landscape evolution for the Quebrada de Purmamarca in NW-Argentina. A thorough mapping of the landforms and the description of the present morphodynamic situation are the starting point for further observation. One of the essential outputs of the study is therefore a geomorphological map of the study area. The results from geomorphological mapping have been supplemented by sedimentological and pedological observations. According to basic morphogenetic principles, the combined results have been ordered chronologically. This way they could finally form the base for landscape evolutionary interpretation. Particularly the discussion of paleoclimatic issues has been emphasized within this study.
A variety of methodical approaches, which all complement each other, have been used. From multipectral LANDSAT TM 5 data, but also from high-resolution CORONA satellite photography and aerial photography, inaccessible parts of the study area could be included in the investigation. In addition these data served overview purposes and has been used as the base for mapping. The main emphasis of this study, however, are the results obtained from several weeks of field work. Aside from geomorphological inquiry and mapping, field work has focused on the description of numerous sedimentological and pedological profiles. The analysis of these profiles was supported by laboratory data from field samples (granulometry, $CaCO_3$ content) but also by a ^{14}C age date. With particular regard to pedological questions, several samples from soil crust were micromorphologically analysed and interpreted.

On the base of these data, several phases of landscape evolution have been reconstructed for the Quebrada de Purmamarca. Depending on their timescale, the different phases show evidence for different controls of landscape change. Landscape evolution of the study area can be traced back approximately into the Miocene, even if geomorphological and sedimentolgical evidence is scarce. During this phase, the Andes were still a landscape of relatively low relief being subject to processes of planation under conditions markedly more humid than today. Within the study area, highly faulted and deformed fanglomerates are the first evidence of a progressing uplift coupled with an increasingly arid climate. As a consequence of continued uplift and alternating phases of erosion and aggradation, large terrace systems have formed. Particularly the youngest terrace level shows well preservation. Against the background of the intense climatic changes characteristic for the Pleistocene, these terraces have been the major focus of this study. They are built up almost entirely from coarse debris-flow sediments which are thought to be the result of a significant drop of the periglacial belt of more than 1,000 meters. This interpretation is confirmed by a variety of relict periglacial landforms like "glatthang" morphology (smooth topography), thick sheets of frost debris and asymmetric valleys. Only a massive drop of temperatures ,typical for most of the last glacial cycle, would have been capable of producing this enormous shift. As the sediment supply from debris production exceeded the transport capacity of the rivers, the fluvial system was virtually choked and an intense phase of aggradation commenced. Aggradation has been interrupted or at least weakened several times as reflected by two marked lacustrine to fluvial layers within the terrace deposits. In this context, particularly the younger layer is of some importance. It announces a shift in morphodynamics and has been dated to approximately 49 ka BP (^{14}C age). Departing from this point, the landscape becomes subject to manifold and relatively frequent environmental changes. These changes cause alternating phases of erosion, stability and accumulation and result in the complex assemblage of relict and present landscape elements.

While the phase of enormous terrace aggradation grades into a phase of dominant alluvial fan activity, the onset of the evident intense incision of the terrace is thought to have occurred relatively late. Particularly on gently dipping terrace segments a well-developed reddish soil has developed. These are interpreted to indicate a phase of increased humidity. Possibly they correlate with a wet phase between 40 ka BP and 25 ka BP, commonly known as the "Minchin" lacustrine event. At many places, this reddish soil is overlain by a markedly cemented sand crust. Based on its evidently well sorting of medium and fine sand, this sand crust has tentatively been interpreted as a fluvio-eolian sediment. Its deposition under very arid and cold climatic conditions has been attributed to the Late Glacial Maximum (LGM). However, it is precisely this sand crust which shows signs of erosion at many places and has not been observed anywhere below the level of the terrace surface. Therefore the onset of severe erosion and incision resulting in the evacuation of enormous quantities of sediment from the study area is assumed to postdate the LGM. Under continued climatic seasonality, it might have been the transition to a wetter phase which is responsible for increased discharge rates. In fact, a phase of relatively humid conditions has been reported from Bolivia and the Cordillera Oriental for late glacial times. Regardless of its timing, the intense incision is likely to have cut down to below the present floodplain evidently causing several mass wasting events in the study area.

Since the early Holocene a number of short-term climatic changes seem to have been responsible for the landscape evolution of the Quebrada de Purmamarca. In this context, more humid phases of pronounced slope smoothing seem to have alternated with semi-arid phases of longer duration. The well-developed, polycyclic calcretes on top of the inactive terraces and alluvial fans give evidence for these changing conditions. Whether this type of soil formation is also characteristic for the present climate could not be decided on the base of the available data. The marked desert pavement on top of most terraces and alluvial fan surfaces rather support very reduced soil formation at present. However, these desert pavements are partly responsible for the intense dynamics along the terrace slopes. By lowering infiltration the pavement serves to concentrate the runoff, leading to badland formation and alluvial fan activity within the terrace deposits. The presently observed floodplain aggradation may be attributed to these processes but considering the severe gullying reaching far into the upper study area, the aggradation might as well reflect a much more general and regional trend. In any case it is particularly this aggradational tendency that has already caused remarkably damage in the past. Therefore it seems doubtful whether the presently undertaken preventive measures make sense, or if they are no more than an additional step within the landscape evolution of the Quebrada de Purmamarca.

RESUMEN

LA QUEBRADA DE PURMAMARCA, JUJUY, NOROESTE ARGENTINO: DESARROLLO DEL PAISAJE Y MORFODINÁMICA EN LOS ANDES SEMIÁRIDOS.

El presente trabajo tuvo como objetivo la reconstrucción de la historia del paisaje en la Quebrada de Purmamarca, en el Noroeste argentino, con especial interés en la discusión sobre interrogantes paleoclimáticos. La base del trabajo la constituye un detallado análisis de las geoformas y el registro de la morfodinámica actual. Parte del producto de la investigación, lo constituye una carta geomorfológica del área de trabajo. Las evidencias geomorfológicas fueron complementadas con observaciones sedimentológicas y pedológicas. Mediante la aplicación de las premisas morfogenéticas, se pudo ordenar los resultados cronológicamente. Solo así fue posible obtener una visión de conjunto sobre la génesis del paisaje.
El trabajo se realizó utilizando diversas metodologías que se complementaron entre si. A través de la interpretación de datos multiespectrales de Landsat TM 5, imágenes satelitales CORONA de alta resolución y fotografías aéreas, pudieron integrarse sectores del área de difícil acceso. Los sensores remotos permitieron, además, contar con una visión general del terreno y constituyeron la base para el mapeo en el campo. El punto central del trabajo son los resultados de varios meses de trabajo en el campo. El interés principal, junto al mapeo de las geoformas, fue el levantamiento de numerosos perfiles sedimentológicos y pedológicos. La interpretación de los perfiles se realizó a través de análisis de laboratorio clásicos (textura, $CaCO_3$), pero también por medio de una datación ^{14}C. En consideración a especial a los interrogantes pedogenéticos, se analizó e interpretó la micromorfología en varios cortes delgados de costras de suelos.

Sobre la base de estos datos, pudieron reconstruirse varias fases de desarrollo del paisaje en la Quebrada de Purmamarca. Los indicadores geomorfológicos y sedimentológicos, si bien incompletos y en parte solo puntuales, aportan información sobre la historia del paisaje a partir ya del Mioceno. En ese momento, los Andes no contaban con el carácter montañoso actual y, bajo un clima marcadamente mas húmedo que el actual, fueron objeto de erosión mantiforme. Las primeras evidencias de orogénesis y un clima progresivamente mas árido, se encuentran en el área de trabajo como fanglomerados muy fracturados y deformados. Como consecuencia del continuo ascenso de los Andes y de la alternancia de fases de acumulación y erosión, se formaron sistemas de terrazas potentes; particularmente el nivel de terraza mas reciente, se ha preservado en diversos sectores. Estas terrazas fueron el aspecto central de la investigación, teniendo en consideración los importantes cambios climáticos ocurridos durante el Pleistoceno. Estas se componen, en casi todo su espesor, de flujos de detritos ("debris flow"), de granulometría muy gruesa, que se interpretan como el resultado de la actividad de un piso periglacial a mas de 1000 metros por debajo del actual. Esta aseveración se fundamenta en un conjunto de formas periglaciales relícticas, tales como "Glatthängen", coberturas detríticas y valles asimétricos. Este descenso del piso periglacial solo pudo haber tenido lugar bajo un descenso generalizado de la temperatura, como por ejemplo el que se comprobó para el último ciclo glacial a nivel mundial. Debido a que la producción de detritos superó en varias magnitudes su capacidad de transporte, el sistema fluvial se redujo notablemente y todo la región estuvo caracterizada por el predominio de la sedimentación. Esta fase de agradación fue interrumpida o debilitada en su intensidad varias veces, lo que se evidencia en dos acumulaciones lacustres. La mas joven de ambas acumulaciones es particularmente relevante, ya que representa la transición hacia condiciones morfodinámicas totalmente diferentes; este depósito fue datado por ^{14}C en alrededor de 49 ka BP. A partir de dicho momento, se suceden numerosas variaciones del clima y fases alternantes de erosión, estabilidad y

sedimentación, que dan lugar a una imagen compleja con elementos del paisaje relícticos y activos.
Mientras que a las enormes acumulaciones que forman las terrazas le sigue una fase de actividad de conos aluviales, hay una serie de evidencias de que la fase de erosión intensiva e incisión es posterior a estas. Especialmente en las superficies planas de las terrazas, se encuentran suelos rojos bien conservados, que se interpretan como evidencia de condiciones mucho más húmedas. Estos resultados pueden correlacionarse posiblemente con una fase húmeda, datada entre aproximadamente 40 ka y 25 ka en la Cordillera Oriental de los Andes Centrales (Fase húmeda "Minchin"). Este suelo rojo aparece cubierto en varios sectores por una costra de arena endurecida. Especialmente debido a la excelente selección de la arena fina y media, que se observa en los cortes delgados, se interpreta preliminarmente que su origen fue fluvio-eólico. Su edad correspondería al último máximo glacial, que tuvo características frías y secas en las regiones vecinas. La costra arenosa se presenta en muchos sectores muy erosionada, pero siempre en la superficie de la terraza. En consecuencia, se interpreta que la fase de erosión intensa es posterior al máximo glacial. Bajo condiciones climáticas mucho más cálidas, en transición hacia una fase húmeda, se produce un aumento de los caudales fluviales, manteniendo la estacionalidad. En los hechos, se habla para el final del Pleistoceno de una fase húmeda en el Altiplano Boliviano así como en la Cordillera Oriental de los Andes Centrales. Independientemente del momento en que ocurrió, la intensa incisión del cauce alcanzó el cauce actual y desencadenó en las vertientes una serie de movimientos en masa.

Desde el Holoceno Inferior parecen haberse sucedido una serie de varias oscilaciones climáticas de corta duración. Las fases húmedas con desarrollo de las vertientes se alternaron con largas fases semiáridas. Debido a esa alternancia, se pueden encontrar varias fases de costras calcáreas bien desarrolladas sobre los conos aluviales inactivos y las superficies de las terrazas. El interrogante sobre si este tipo de desarrollo de suelos también es característico de la situación climática actual, no puede ser respondido aún con los datos disponibles. La cobertura de muchas superficie con pavimentos del desierto indican mas bien un desarrollo de suelos muy limitado. A pesar de ello, estos pavimentos son corresponsables de la alta actividad geomorfológica en las terrazas, ya que disminuyen la infiltración y conducen a la concentración del escurrimiento y con ello a la formación de Badlands en terrazas y conos aluviales. La tendencia actual a la acumulación en los cauces puede relacionarse al importante aporte de sedimentos de estos sectores de Badlands, pero, también, teniendo en cuenta que hay cárcavas que alcanzan los sectores altos de la cuenca, deben ser también consecuencia de una tendencia general regional. En todos los casos, es esta importante dinámica aluvial y el transporte de sedimentos el que en el pasado ocasionara numerosos daños en la región. Por ello es discutible si las medidas de prevención puntuales que deben evitar los graves daños actuales, tienen sentido a largo plazo o solo representan el último paso de la génesis del paisaje en la Quebrada de Purmamarca.

ZUSAMMENFASSUNG

DIE QUEBRADA DE PURMAMARCA, JUJUY, NW-ARGENTINIEN: LANDSCHAFSENTWICKLUNG UND MORPHODYNAMIK IN DEN SEMI-ARIDEN ANDEN.

Die vorliegende Arbeit hatte zum Ziel, die Landschaftsgeschichte der Quebrada de Purmamarca in NW-Argentinien zu rekonstruieren. Eine eingehende Analyse der einzelnen Landformen und die Aufnahme der aktuellen morphodynamischen Situation dienten dabei als Arbeitsgrundlage. Eine geomorphologische Karte des Arbeitsgebietes ist somit eines der Produkte der Untersuchung. Ergänzt wurden diese geomorphologischen Befunde vor allem durch sedimentologische sowie bodenkundliche Beobachtungen. Unter Anwendung grundlegender morphogenetischer Prinzipien wurden diese Ergebnisse nun chronologisch geordnet. Nur so konnten sie abschließend in ihrer Gesamtheit der landschaftsgenetischen Interpretation dienen. Innerhalb dieser Arbeit stand dabei insbesondere die Diskussion paläoklimatischer Fragestellungen im Vordergrund.
Insgesamt gesehen, stützt sich die Arbeit auf eine Vielzahl methodischer Ansätze, die sich gegenseitig ergänzen. Durch die Auswertung multispektraler LANDSAT TM 5 Daten, hochauflösender CORONA-Satellitenbilder und Luftbilder konnten auch schwer zugängliche Gegenden in die Untersuchung eingebunden werden. Außerdem dienten diese Fernerkundungsdaten dem Überblick und waren Kartiergrundlage im Gelände. Eigentlicher Schwerpunkt der Arbeit allerdings sind die Ergebnisse mehrmonatiger Feldforschung. Besonderes Augenmerk wurde neben der Kartierung geomorphologischer Formen auf die Aufnahme zahlreicher sedimentologischer und bodenkundlicher Profile gelegt. Die Auswertung aller Profile wurde dabei durch die üblichen sedimentologischen Laboranalysen (Korngrößen, $CaCO_3$), aber auch durch eine ^{14}C-Altersdatierung unterstützt. Insbesondere im Hinblick auf pedogenetische Fragen wurden schließlich mehrere Dünnschliffe von Bodenkrusten mikromorphologisch untersucht und gedeutet.

Auf der Basis dieser Daten lassen sich für die Quebrada de Purmamarca mehrere Phasen der Landschaftsentwicklung zurückverfolgen, die je nach abgedeckten Zeitskalen unterschiedliche Interpretationen im Hinblick auf landschaftsgeschichtliche Einflussgrößen erlauben. Wenn auch nur unvollständig und teilweise punktuell, gehen geomorphologische wie auch sedimentologische Nachweise der Landschaftsentwicklung bis etwa ins Miozän zurück. Zu dieser Zeit hatten die Anden noch nicht ihren heutigen Hochgebirgscharakter und waren in deutlich humiderem Klima einer flächenhafteren Abtragung unterlegen. Erste Hinweise auf eine voranschreitende Gebirgsbildung und ein sich zunehmend aridisierendes Klima finden sich im Untersuchungsgebiet in der Form von intensiv gestörten und deformierten Fanglomeraten. Als Konsequenz der andauernder Gebirgsbildung und alternierenden Phasen von Aufschüttung und Erosion entstehen mächtige Terrassensysteme. Vor allem das jüngste Terrassenniveau ist an vielen Stellen erhalten. Vor dem Hintergrund der nachweislich starken Klimaschwankungen während des Pleistozäns, wurden diese Terrassen in den Mittelpunkt der Arbeit gestellt. In nahezu ihrer gesamten Mächtigkeit bestehen sie aus enorm grobkörnigen „debris-flow" Sedimenten und werden als Resultat einer bis über 1,000 Meter tieferen Periglazialstufe gedeutet. Untermauert wird diese Interpretation durch eine Reihe von reliktisch erhaltenen periglazialen Formen wie Glatthängen, Schuttdecken und assymetrischen Tälchen. Nur eine massive Temperaturabsenkung, wie sie für den Verlauf des letzten Glazials weltweit belegt ist, kann diese Absenkung hervorgerufen haben. Da das Schuttaufkommen die Transportkapazität um ein Vielfaches überschritt, wurde das fluviale System praktisch lahm gelegt, und die gesamte Region war gekennzeichnet von Sedimentation. Diese Phase der Sedimentanhäufung war mehrfach in ihrer Intensität unterbrochen oder abgeschwächt, was sich anhand

zweier auffälliger Lagen lakustriner Sedimente nachweisen lässt. Vor allem die jüngere von beiden ist in diesem Zusammenhang von Wichtigkeit, da sie den Übergang zu deutlich veränderter Morphodynamik einleitet und auf etwa 49 ka BP (^{14}C-Alter) datiert worden ist. Ab diesem Zeitpunkt ist das Untersuchungsgebiet einer Vielzahl von Klimaschwankungen ausgesetzt und die sich abwechselnden Phasen von Erosion, Stabilität und Sedimentation ergeben ein komplexes Bild aus reliktischen und aktiven Landschaftselementen.
Während sich an die enorme Terrassenschüttung eine Phase von Schwemmfächeraktivität anschließt, gibt es eine Reihe von Hinweisen, dass die Phase der augenscheinlich intensiven Erosion und Einschneidung der Terrassen erst relativ spät eingesetzt hat. Insbesondere auf den flacheren Terrassenoberflächen finden sich recht gut erhaltene rote Böden, die als Zeichen einer deutlich erhöhten Humidität gedeutet wurden. Möglicherweise können diese Ergebnisse mit einer Feuchtphase korreliert werden, die sich für den Zeitraum zwischen etwa 40 ka BP und 25 ka BP im der gesamten Ostkordillere der zentralen Anden belegen lässt („Minchin“ Feuchtphase). Überlagert wird dieser rote Boden vielerorts von einer auffällig verhärteten Sandkruste. Vor allem aufgrund der im Dünnschliff erkennbaren sehr guten Sortierung dieser Kruste aus Fein- und Mittelsand wurde sie vorsichtig als fluvio-äolische Bildung interpretiert. Damit fällt sie zeitlich aller Wahrscheinlichkeit ins Hochglazial, das auch in angrenzenden Regionen besonders kalt und trocken gewesen sein muss. Schließlich ist es aber gerade diese Sandkruste, die sich an vielen Stellen deutlich erodiert zeigt und an keiner Stelle unterhalb der Terrassenoberfäche vorkommt. Folglich wird für das Einsetzen der Phase intensivster Erosion und Sedimentausräumung ein Zeitpunkt nach dem Hochglazial angenommen. Dabei könnte, bei sich deutlich erwärmendem Klima, der Übergang in eine Feuchtphase die Abflussmengen unter Beibehaltung der Saisonalität deutlich erhöht haben. In der Tat wird im ausgehenden Pleistozän von einer Feuchtphase auf dem bolivianischen Altiplano sowie in der zentralandinen Ostkordillere berichtet. Unabhängig vom ihrem Zeitpunkt reicht diese intensive Einschneidung unter den heutigen Talboden hinab und löst an den Hängen eine Reihe von Massenbewegungen aus.

Seit dem frühen Holozän scheinen sich mehrere verschiedene, relative kurze Klimaschwankungen auf die Landschaft in der Quebrada de Purmamarca ausgewirkt zu haben. Dabei scheinen feuchtere Phasen mit betonter Hangentwicklung mit längeren semi-arid Phasen zu wechseln. Durch diesen Wechsel lassen sich die recht gut entwickelten, mehrphasigen Kalkrusten auf den inaktiven Schwemmfächern und Terrassenoberflächen erklären. Ob diese Art der Bodenentwicklung allerdings auch charakteristisch ist für die aktuelle Klimasituation im Untersuchungsgebiet, lässt sich aufgrund der vorhanden Daten nicht genau erschließen. Die auffällige Bedeckung vieler Flächen mit einem Wüstenpflaster deuten eher auf eine sehr eingeschränkte Bodenbildung hin. Dennoch sind diese Pflaster mitverantwortlich für die hohe Aktivität an Terrassenabhängen und -kanten, da sie die Infiltration erniedrigen und somit zur Abflusskonzentration und zur Badlandbildung in den Terrassen- und Schwemmfächersedimenten führen. Die aktuell beobachtete Tendenz der Aufschüttung des Talbodens mag zum einen in einer erhöhten Sedimentzufuhr aus diesen Badlandbereichen begründet sein, muss aber angesichts der weit ins obere Einzugsgebiet reichenden Gullies auch als eine generelle regionale Tendenz gedeutet werden. In jedem Fall ist es genau diese hohe Dynamik der Aufschüttung und Sedimentbewegung, die in der Vergangenheit schon für vielerlei Schäden in der Region gesorgt hat. Somit ist fraglich, ob die sehr lokalen präventiven Maßnahmen, die aktuell schlimmere Schäden verhindern sollen, überhaupt eine längerfristigen Sinn machen, oder nur den jüngsten Schritt der Landschaftsgenese der Quebrada de Purmamarca darstellen.

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MAPS

IGM (1986): Carta Topográfica de la República Argentina 1:250.000: Hoja 2366-IV Ciudad del Libertador San Martín.

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DATA

GASATACAMA (1997): Aerial-Photography ~1:26,000
(Frames 11472-24-5 to 7, 11472-24-10 to 12, 11472-27-3 to 8).

IGM (1966) : Aerial-Photography 1:65,000
(Frames 2365-307-16 to 19, 2365-308-15/16, 2365-311-28/29).

LANDSAT 5 TM SATELLITE-DATA
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APPENDIX

Locations of sedimentological and pedological profiles and samples

List of GPS points from the study area

Data from granulometric and $CaCO_3$ analysis

Results from ^{14}C-dating

Geomorphological map of the Quebrada de Purmamarca (~1:85,000)

Geomorphological map of the Qbd. de Purmamarca (~1:85,000) - transparent

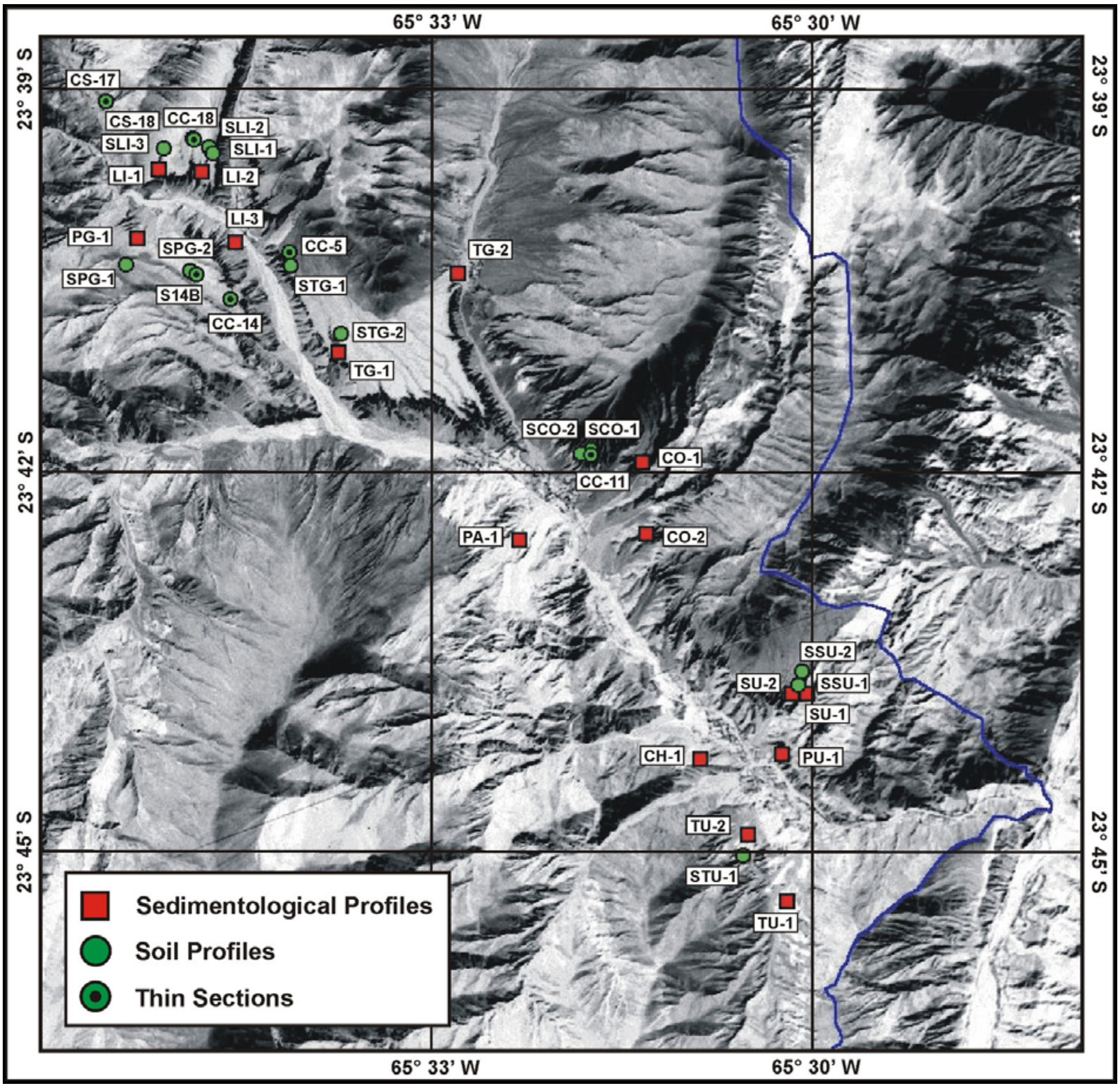

Fig. 205: Locations of sedimentological profiles, soil profiles and samples for thin sections within the lower study area..

Name	LAT	LONG	Altidude
CH 1	-23,7365	-65,5195	2522,5
CH 2	-23,7359	-65,5147	x
CH 3	-23,7346	-65,5182	2479
CH 4	-23,7387	-65,5172	2569,5
CO 1	-23,7004	-65,5216	2795
CO 2	-23,6986	-65,5182	2933
CO 3	-23,6974	-65,5285	2872
EG 1	-23,6099	-65,5333	3486,5
EG 2	-23,6129	-65,5246	3385
EG 3	-23,6739	-65,5459	2902,5
EG 4	-23,6861	-65,5422	2780
LI 1	-23,6751	-65,5738	x
LI 2	-23,6658	-65,5675	x
LI 3	-23,6556	-65,5830	x
LI 4	-23,6555	-65,5827	x
LI 5	-23,6670	-65,5780	x
LI 6	-23,6627	-65,5774	3170
LI 7	-23,6639	-65,5803	x
LI 8	-23,6695	-65,5707	x
LI 9	-23,6704	-65,5735	x
LI 10	-23,6706	-65,5735	x
LI 11	-23,6687	-65,5750	x
LI 12	-23,6504	-65,5773	3210
LI 13	-23,6532	-65,5800	x
LI 14	-23,6686	-65,5749	x
LI 15	-23,6544	-65,5773	x
LI 16	-23,6681	-65,5736	x
LI 17	-23,6573	-65,5780	3190
LI 18	-23,6610	-65,5858	3153
LI 19	-23,6625	-65,5840	3076
LI 20	-23,6578	-65,5839	x
LI 21	-23,6562	-65,5801	x
LI 22	-23,6577	-65,5772	3160
LI 23	-23,6624	-65,5867	3153
LI 24	-23,6617	-65,5850	3150
PA 1	-23,7003	-65,5347	2670
PA 2	-23,7082	-65,5373	2696
PA 3	-23,7122	-65,5534	3126
PA 4	-23,7105	-65,5390	2833,5
PA 5	-23,7101	-65,5372	2808,5
PG 1	-23,6817	-65,5749	2985
PG 2	-23,6767	-65,5768	x
PG 3	-23,6852	-65,5669	x
PG 4	-23,6767	-65,5766	x
PG 5	-23,6842	-65,5696	2879
PG 6	-23,6776	-65,5766	x
PG 7	-23,6751	-65,5771	x
PG 8	-23,6760	-65,5766	3156
PG 9	-23,6806	-65,5740	3100
PG 10	-23,6677	-65,5873	3120
PG 11	-23,6759	-65,5744	x
PG 12	-23,6765	-65,5739	3079
PG 13	-23,6728	-65,5889	3280
PG 14	-23,6759	-65,5750	x
PG 15	-23,6702	-65,5932	3262
PG 16	-23,6756	-65,5819	3215
PG 17	-23,6740	-65,5801	3186,5
PG 18	-23,6664	-65,5893	3190
PO 1	-23,6748	-65,6108	3475
PU 1	-23,7451	-65,5010	2366
PU 2	-23,7456	-65,4988	2365
PU 3	-23,7387	-65,5020	2555
QL 1	-23,6370	-65,5721	x
QL 2	-23,6499	-65,5728	x
QL 3	-23,6437	-65,5719	x
QL 4	-23,6479	-65,5719	3128
QL 5	-23,6451	-65,5721	x
QL 6	-23,6352	-65,5720	3240
QS 1	-23,7121	-65,5252	2616
SE 1	-23,6603	-65,5940	3260
SE 2	-23,6595	-65,5931	3230
SE 3	-23,6428	-65,5959	3340
SE 4	-23,6514	-65,5915	3267
SE 5	-23,6365	-65,5989	3438
SE 6	-23,6528	-65,5931	3192
SU 1	-23,7322	-65,5035	2523
SU 2	-23,7374	-65,5078	2429,5
SU 3	-23,7322	-65,5010	2614
SU 4	-23,7286	-65,4982	2699,5
SU 5	-23,7280	-65,5015	2686
TG 1	-23,6885	-65,5613	2850
TG 2	-23,6663	-65,5666	3159,5
TG 3	-23,6796	-65,5652	3020
TG 4	-23,6773	-65,5557	3063
TG 5	-23,6875	-65,5613	2850
TG 6	-23,6922	-65,5603	2780
TG 7	-23,6897	-65,5593	2940
TG 8	-23,6712	-65,5677	3052
TG 9	-23,6731	-65,5676	3040
TU 1	-23,7477	-65,5074	2423
TU 2	-23,7572	-65,5034	x
TU 3	-23,7566	-65,5037	2502
TU 4	-23,7540	-65,5045	2471
TU 5	-23,7506	-65,5085	2588,5
TU 6	-23,7483	-65,5087	2529,5

CH Chalala, **CO** Qda. del Cobre, **EG** Estancia Grande **LI** Lipán, **PA** Patacál, **PG** Potrero Grande, **PO** Qda. de Potrerillos, **PU** Purmamarca, **QL** Qda. de Lipán **QS** Quisciri, **SE** Sepulturas, **SU** Qda. de Sunchoguaico, **TG** Terraza Grande, **TU** Qda. de Tumbaya

Table 6: List of GPS points (used for georeferencing LANDSAT 5 TM and CORONA satellite data).

	cS	mS	fS	cU	mU	fU	T	S total	U total	Max. clast	Ø Layer
	[W.-%]									[cm]	[cm]
LI 1	14,30	13,47	21,40	15,67	11,94	7,16	15,57	49,17	34,77	40	-
LI 2	13,99	11,09	21,57	19,88	13,77	8,14	13,67	46,65	41,79	70	-
LI 3	-	-	-	-	-	-	-	-	-	-	-
LI 4	13,28	11,02	20,71	19,06	13,25	8,63	14,49	45,01	40,94	40	-
LI 5	12,02	10,60	23,97	21,40	13,40	7,38	12,16	46,59	42,18	60	-
LI 6	12,32	9,46	22,42	22,41	13,49	8,17	13,14	44,20	44,07	30	-
LI 7	13,88	10,13	23,11	19,25	13,39	7,06	12,97	47,12	39,70	50	-
LI 8	13,15	9,64	24,46	21,12	12,91	7,51	12,15	47,25	41,54	100	-
LI 9	26,45	14,06	22,40	11,67	8,35	5,22	8,26	62,91	25,24	80	-
LI 10	14,91	11,52	20,88	17,12	14,35	8,04	13,04	47,31	39,51	100	-
LI 11	15,41	10,99	20,13	18,52	13,63	8,05	12,47	46,53	40,20	80	-
LI 12	14,25	14,09	24,25	17,26	12,30	7,06	10,15	52,59	36,62	60	-
LI 13	25,71	20,04	17,52	12,15	9,23	6,26	9,03	63,27	27,64	50	-
LI 14	23,76	16,37	18,11	12,91	10,19	6,83	10,29	58,24	29,93	30	-
LI 15	27,20	17,22	17,92	13,14	8,97	5,46	9,14	62,34	27,57	30	-
TG 1	20,32	17,08	22,46	13,48	9,88	5,70	10,19	73,34	25,77	40	150
TG 2	38,02	34,98	14,13	5,22	2,44	1,76	3,32	92,35	7,52	60	100
TG 3	12,94	11,69	18,99	16,44	13,30	4,26	21,37	60,06	38,93	45	150
TG 4	11,14	13,87	21,55	18,81	14,16	8,29	14,42	65,37	36,87	100	180
TG 5	21,72	25,35	28,71	6,72	9,12	3,49	4,36	82,50	16,97	80	100
TG 6	15,08	11,31	19,43	16,88	13,42	8,82	15,12	62,70	37,36	20	50
TG 7	10,96	12,09	22,98	19,40	13,16	7,74	15,08	65,43	35,98	50	50
TG 8	18,22	16,03	29,65	17,11	7,09	4,75	8,49	81,01	20,33	50	180
TG 9	18,69	11,26	20,39	18,33	13,40	7,41	11,03	68,67	31,84	50	100
TG 10	15,25	9,08	19,85	19,92	13,87	9,48	13,52	64,10	36,87	45	100
TG 11	9,13	8,83	24,02	22,77	13,81	7,56	15,39	64,75	36,76	90	250
TG 12	10,46	8,06	21,69	21,08	15,43	9,15	15,19	61,29	39,77	90	200
TG 13	11,15	17,65	30,40	17,59	11,07	6,93	5,36	76,79	23,36	60	150
TG 14	16,97	10,99	19,64	17,92	13,22	8,52	12,91	65,52	34,65	50	100
TG 15	11,67	11,18	24,08	20,89	14,73	9,26	7,79	67,82	31,78	40	100
TG 16	33,27	18,60	17,40	11,42	6,95	4,06	7,51	80,69	18,52	30	150

Table 7: Results from grainsize analysis of matrix samples from LI-2 and TG-1.

	cS	mS	fS	cU	mU	fU	T	$CaCO_3$
	[W.-%]							[%]
STG-1A	16,83	12,23	25,51	7,19	7,48	8,15	22,63	26,03
STG-1B	3,78	15,84	63,71	6,92	0,78	1,20	7,77	0,85
STG-1C	9,46	18,43	60,27	0,69	3,09	0,86	7,21	0,81
STG-2A	4,77	10,31	42,13	13,06	5,66	9,46	14,61	0,21
STG-2B	1,24	5,99	27,65	7,90	16,55	15,37	25,30	0,94
SCO-1A	2,24	3,67	24,86	8,07	16,26	15,12	29,79	17,27
SCO-1B	5,15	18,39	36,25	8,04	4,80	8,04	19,32	24,59
SCO-2A	2,93	5,20	26,04	11,19	18,76	12,51	23,37	22,35
SCO-2B	15,13	11,85	20,80	13,11	16,28	6,01	16,82	43,10
SSU-1A	5,39	12,54	50,39	15,67	4,94	0,85	10,22	15,45
SSU-1B	8,78	11,52	41,95	0,24	10,21	0,24	27,54	17,27
SSU-2	14,82	13,71	54,35	1,99	4,20	1,65	9,28	4,29
SPG-1A	4,61	10,78	57,75	11,37	6,51	2,80	6,17	0,56
SPG-1B	2,33	5,82	21,29	3,40	14,00	16,55	36,60	0,86
SPG-2A	4,93	8,77	43,50	10,85	6,37	8,26	17,32	15,27
SPG-2B	19,28	15,36	31,42	5,85	7,52	5,43	15,14	59,96
STU-1	17,84	16,42	25,66	4,25	7,50	8,06	20,26	31,70
SLI-2A	3,29	23,15	58,56	4,54	2,01	1,37	7,08	0,73
SLI-2B	6,49	18,13	52,89	4,54	5,79	1,14	11,02	0,68

Table 8: Results from grainsize and $CaCO_3$ analysis of the soil samples.

LI1		LI2		TG1	
Direction [°]	Dip [°]	Direction [°]	Dip [°]	Direction [°]	Dip [°]
170	5	200	2	160	5
170	2	200	6	120	2,5
180	2,5	210	4	210	2,5
180	4	170	4	240	10
235	5	160	4	170	6
230	5	210	8		
		160	5		
		175	5		

CO-2		SU-1 / SU-2	
Direction [°]	Dip [°]	Direction [°]	Dip [°]
320	2	140	2
240	2	95	3
300	2	150	1,5
0	2	150	2,5
		90	2,5
		140	2,5

Table 9: Paleoflow measurements in lithofacies D and F in selected sedimentological profiles.

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Kiel, den 29. April 2002

Datierungsergebnisse der Probe KIA 17090

Sehr geehrter Herr Richter,

anbei übersende ich Ihnen die Ergebnisse der Datierung der oben genannten Probe.

Die Probe wurde unter dem Mikroskop auf Verunreinigungen kontrolliert. Die Probe selbst besteht aus Holzkohleresten mit Sand / Silt und enthält leider auch einen hohen Anteil an Fremdverunreinigungen in Form von Textilfasern. Zur weiteren Analyse wurde eine Teilprobe mit sowenig wie möglich Fasern entnommen. Diese wurde dann mit 1 % HCl, 1 % NaOH und wieder 1 % HCl bei 60 °C extrahiert. Die Verbrennung erfolgte bei 900 °C in einer mit CuO und Silberwolle gefüllten Quarzampulle. Das entstandene CO_2 wurde dann mit H_2 bei 600 °C über einem Eisen-Katalysator zu Graphit reduziert und das Eisen-Graphit-Gemisch in einen Probenhalter für die AMS-Messung gepreßt.

Die ^{14}C-Konzentration der Probe ergibt sich aus dem Vergleich der simultan ermittelten ^{14}C, ^{13}C und ^{12}C Gehalte mit denen des CO_2-Meßstandards (Oxalsäure II) sowie einer geeigneten Nulleffekt-Probe. Das konventionelle ^{14}C-Alter berechnet sich anschließend nach einer Korrektur auf Isotopenfraktionierung anhand des ^{13}C/^{12}C-Verhältnisses [Stuiver and Polach, Radiocarbon, **19**/3 (1977), 355]. Dieser δ^{13}C-Wert enthält auch die Effekte der während der Graphitisierung und in der AMS-Anlage auftretenden Isotopenfraktionierung und ist deshalb nicht direkt vergleichbar mit δ^{13}C-Werten, die in einem CO_2-Massenspektrometer gemessen werden. Die Unsicherheit im ^{14}C-Ergebnis berücksichtigt Zählstatistik, Stabilität der AMS-Anlage und Unsicherheit im subtrahierten Nulleffekt. Für die ersten beiden haben wir die Zählstatistik und die beobachtete Streuung der Meßintervalle verglichen und den größeren Wert verwendet.

Die Probe hat mehr als die für eine präzise Datierung empfohlene Mindestmenge von ca. 1 mg Kohlenstoff und damit ausreichend Probenstrom in der AMS-Anlage ergeben. Der δ^{13}C-Wert liegt im Normalbereich für organische Proben. Die Ergebnisse sind insofern zuverlässig.

Mit einem Radiokarbonalter von 49 550±1700 Jahren BP wäre die Probe wesentlich älter als von Ihnen vermutet. Die große Differenz zwischen dem erwarteten Altersbereich von 10 000 bis 20 000 Jahren und dem gemessenen Alter lässt sich allein durch die Verunreinigung mit älterem, ^{14}C-freien organischen Material der in der Probe noch enthaltenen Textilfasern (petrochemische Produkten) nicht erklären (erfordert > 95 % Verunreinigung). Das Probenalter dürfte tatsächlich > 20 000 Jahre sein, wenn Sie aus dem siltigen Sediment organische Reste gesammelt haben die meistens zur Zeit der Sedimentation alt waren und umgelagert wurden. Das organische Mischalter wäre dann viel älter als der Zeitpunkt der Sedimentation. Wir hatten letzlich einen ähnlichen Fall mit aus sehr organisch-armem Löss (kalt-arides Klima) angereicherten organischen Resten, die viel Älter datierten als zwischengeschaltete organisch-reiche Schichten. Könnte so etwas bei Ihnen zutreffen?

Wenn Sie zu diesen Datierungen Fragen haben, stehe ich gerne zu Ihrer Verfügung.

Mit freundlichen Grüßen

(P.M. Grootes)

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† „PMC (korrigiert)“ bezeichnet den prozentualen Anteil an modernem (1950) Kohlenstoff, korrigiert auf Massenfraktionierung mittels der ^{13}C Messung.
‡ Bitte beachten Sie, dass der $\delta^{13}C$ Wert Fraktionierungen in der Probenaufbereitung sowie während der AMS Messung beinhaltet und daher nicht mit einer massenspektrometrischen Messung verglichen werden kann.

Datierungsergebnisse der Probe KIA 17090

KIA17090 **PU-L12U**

Kohlereste, Purmamarca / NW-Argentinien, Entnahmetiefe: 20 cm

Fraktion	PMC (korrigiert)†	Radiokarbonalter	$\delta^{13}C$(‰)‡
S Fussel, Holzkohle, Laugenrückstand, 5.1 mg C	0.21 ± 0.04	49550 + 1690 / -1400 BP	-22.18 ± 0.12